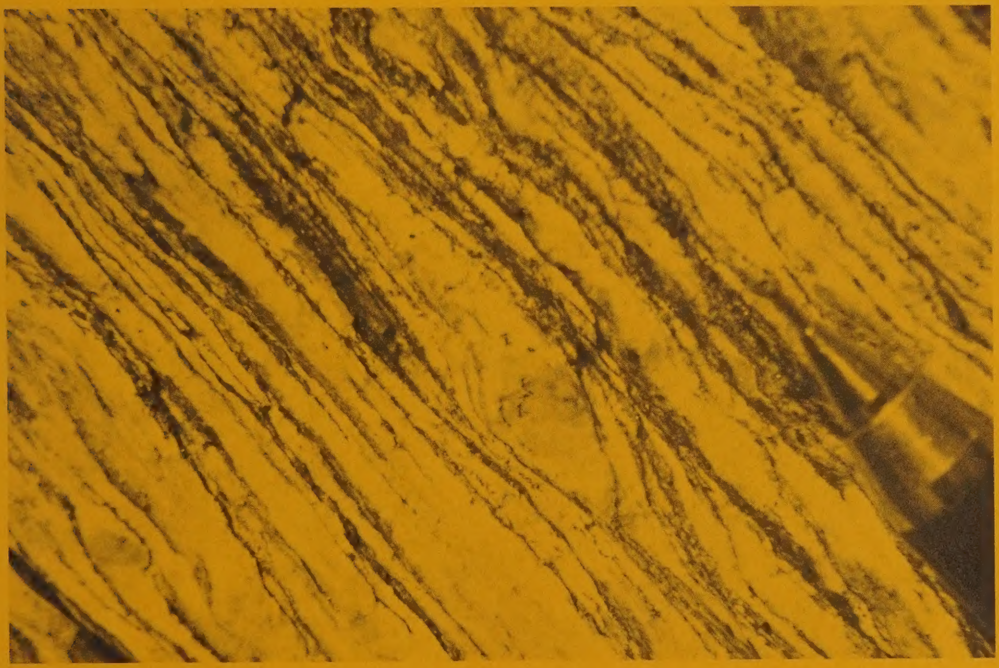
BEDROCK GEOLOGY OF THE NORTHERN PART OF THE LINCOLN MASSIF CENTRAL VERMONT

by

Vincent DelloRusso and Rolfe S. Stanley

University of Vermont, Burlington, Vermont, 05405



Vermont Geological Survey

Charles A. Ratte, State Geologist

Special Bulletin No.8, 1986

Cover Photograph. Pervasive mylonitic schistosity in a mylonite zone at location Q12 (Pl. 2) viewed looking north at rare, relict feldspar porphyroclasts with elongate recrystallized tails. The consistent asymmetry of these porphyroclast tails indicates an east-over-west sense of shear across the zone (left-lateral in photo). Note the planarity of the mylonitic fabric. In thin section, quartz and feldspar are strongly recrystallized and exhibit features of dynamic recrystallization. Pencil tip for scale.

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FIGURES

Four Plates with their own sequence of figures also accompany this text. Figures on these plates are referred to in the text as Pl. #, Fig. #.

ABSTRACT

Detailed mapping of the northern part of the Lincoln massif has delineated 12 lithic units within the Middle Proterozoic Mt. Holly Complex. Gneisses of the western part of the Lincoln massif (WLM) represent a layered sequence of pre-Grenvillian metasedimentary rocks composed of granitic gneiss, augen gneiss, biotite gneiss, and quartzite. In contrast, the granitic gneiss of the eastern part of the Lincoln massif (ELM) is orthogneiss and is overlain by a series of metasedimentary rocks consisting of quartzite and tourmaline-chloritoid schist along the western margin, and augen gneiss and quartzite along the eastern margin of the ELM. Relict Grenvillian metamorphic minerals indicate temperature-pressure conditions of the epidote-amphibolite facies/garnet zone. The granitic gneiss of the ELM is intruded by post-Grenvillian mafic dikes that may be related to the Late Proterozoic rift history of the basement.

Cover rocks are separated into an Eastern Sequence and a Western Sequence. The Western Sequence contains, from oldest to youngest, conglomerate and metawacke (Pinnacle Formation), phyllite (Fairfield Pond Formation), and quartzite (Cheshire Formation) documenting sedimentation along the western margin of Iapetus from the rift stage through the rift-drift transition during Late Proterozoic/Early Cambrian time. The Eastern Sequence contains more distal facies of the pre-Cheshire rift-clastic rocks represented by metawacke and schist interlayered with mafic schist (Hoosac and Underhill Formations).

Major folds and faults in the study area reveal the progressive Taconian evolution of the Lincoln massif as a faulted series of folds within an allochthonous basement slice. Paragneisses of the WLM are retrograded to lower greenschist facies and deformed into Paleozoic, upright, open folds associated with the development of the Ripton anticline. The basement-cover unconformity of the WLM is passively folded. A poorly exposed, narrow, north-plunging syncline separates the WLM from the ELM, with no evidence for major faults within this structure. In contrast, orthogneiss of the ELM is retrograded to lower greenschist facies in the west and epidote-amphibolite facies in the east and exhibits extensive Paleozoic faulting. Thrust zones become increasingly more prevalent eastward from a zone of high-angle reverse faults along the western margin of the ELM. These reverse faults develop from parasitic folds on the western, upright limb of the Lincoln anticline. A N-NE trending zone of extensive Paleozoic mylonites, herein named the Cobb Hill thrust zone, forms a strong topographic lineament within the ELM and displays east-over-west displacement. This thrust zone may correlate with regional thrust zones recognized in the Green Mountain massif of southern Vermont, and may represent the southern extension of the Hinesburg fault zone of northern Vermont. Slivers of mylonitic gneiss delineate major Paleozoic thrusts in the eastern-cover sequence, near the basement-cover boundary. This zone may correlate with the Middlefield thrust zone in Massachusetts and thus define a zone of detachment between the basement massifs and the eastern cover rocks.

INTRODUCTION

Recent bedrock studies in western Massachusetts and Vermont have provided important new data on the geological evolution of this part of the Appalachian orogen (Ratcliffe, 1975, 1982; Stanley, 1978; Ratcliffe and Hatch, 1979; Stanley and Ratcliffe, 1980, 1982; Stanley and Roy, 1982; Stanley and others, 1982, 1984; Doolan and others, 1982; Tauvers, 1982a, 1982b; DiPietro, 1983a, 1983b; Dorsey and others, 1983). Much of this information has been incorporated into a synthesis of the Taconian evolution of western New England by Stanley and Ratcliffe (1983, 1985). The results of these studies have lead to interpretations of the geologic history that contrast sharply with interpretations that evolved prior to the influence of plate tectonic theory.

Detailed mapping by Stanley and others (1982, 1984, 1985) and Doolan and others (1982) in the slope-rise rocks of northern Vermont has revealed several major thrust zones within what was once considered a simple, east dipping, homoclinal sequence (Fig. 1). Numerous studies in western Massachusetts have also shown that much of the stratigraphy east of the Berkshire massif is allochthonous (Ratcliffe, 1975; Stanley, 1978; Ratcliffe and Hatch, 1979; Stanley and Ratcliffe, 1980, 1982). Thrust zones mapped along the eastern margin of the Berkshire massif are interpreted to be continuous with those mapped in the slope-rise rocks of northern Vermont (Stanley and Ratcliffe, 1985).

Recent attention has been focused on the extent of thrust development within the external, Proterozoic basement massifs of New England (Ratcliffe, 1982; Ando and others, 1983, 1984). Ratcliffe (1975, 1983 in Zen and others, 1983) has demonstrated that extensive thrust faults relating to the Taconian and Acadian orogenies occur within the Berkshire massif and that this massif is an allochthonous slice of North American (Grenvillian) basement.

Prior to this study, there was little information bearing on the extent of Paleozoic thrust development to the north in the Green Mountain and Lincoln massifs of Vermont. Ongoing work in the Green Mountain massif has recently defined major Paleozoic thrust zones (Karabinos and Thompson, 1984; Ratcliffe and Burton pers. comm., 1986). The focus of this study is on the depositional history, structural evolution, and metamorphic chronology of the basement and cover rocks that comprise the northern part of the Lincoln massif in central Vermont. In light of the deformational history of the Berkshire massif, particular emphasis is given to the recognition and evaluation of Paleozoic fault zones within the Lincoln massif.

REGIONAL GEOLOGIC SETTING

The Lincoln massif represents the northernmost exposure of Proterozoic basement rocks that core the Green Mountain anticlinorium in Vermont. These rocks which separate contrasting Paleozoic stratigraphic sequences occur along the transition zone between the Taconian foreland to the west and the hinterland to the east. On the west, the basement is unconformably overlain by a sequence of schist and metawacke deposited during the Late Proterozoic rifting of Iapetus (Tauvers, 1982a, 1982b). These rocks are overlain by a thin sequence of laminated phyllite and schist that grades upward into Lower

Cambrian quartzite representing the transition from the rift to the drift stage. Conformably overlying the quartzite to the west is a thick sequence of interlayered carbonate and siliciclastic rocks which comprise the ancient carbonate shelf that developed from Lower Cambrian to Lower Ordovician time. To the east of the Lincoln massif, the basement is overlain by a severely deformed sequence of metawacke, mafic schist, and aluminous, albitic, and carbonaceous schists that are considered to be an eastern facies of the western rift-facies rocks (Stanley and others, 1985).

Platform rocks display well-developed pressure solution cleavage and have an abundance of minor, bed-parallel thrust faults and small-scale duplex structures which are typical of the Taconian foreland (Leonard, 1985; Stanley and others, 1987). The pre-Silurian slope-rise rocks of eastern Vermont are strongly deformed and contain multiple generations of foliation and fault development which are typical of the hinterland (Stanley and others, 1985).

Several major thrust zones are recognized throughout the region. The Champlain thrust, west of the Lincoln massif, is a regionally continuous thrust zone along which the platform and hinterland rocks were transported westward (Fig. 1; Coney and others, 1972; Stanley and Ratcliffe, 1985). North of the Lincoln massif, the Hinesburg thrust zone separates the eastern, slope-rise rocks from the platform sequence of western Vermont (Fig. 1; Dorsey and others, 1983). The pre-Silurian rocks of eastern Vermont have recently been interpreted to be a complex series of thrust slices separated by regionally continuous thrust zones which developed during the Taconian and Acadian orogenies (Stanley and Ratcliffe, 1985).

LOCATION OF THE STUDY AREA

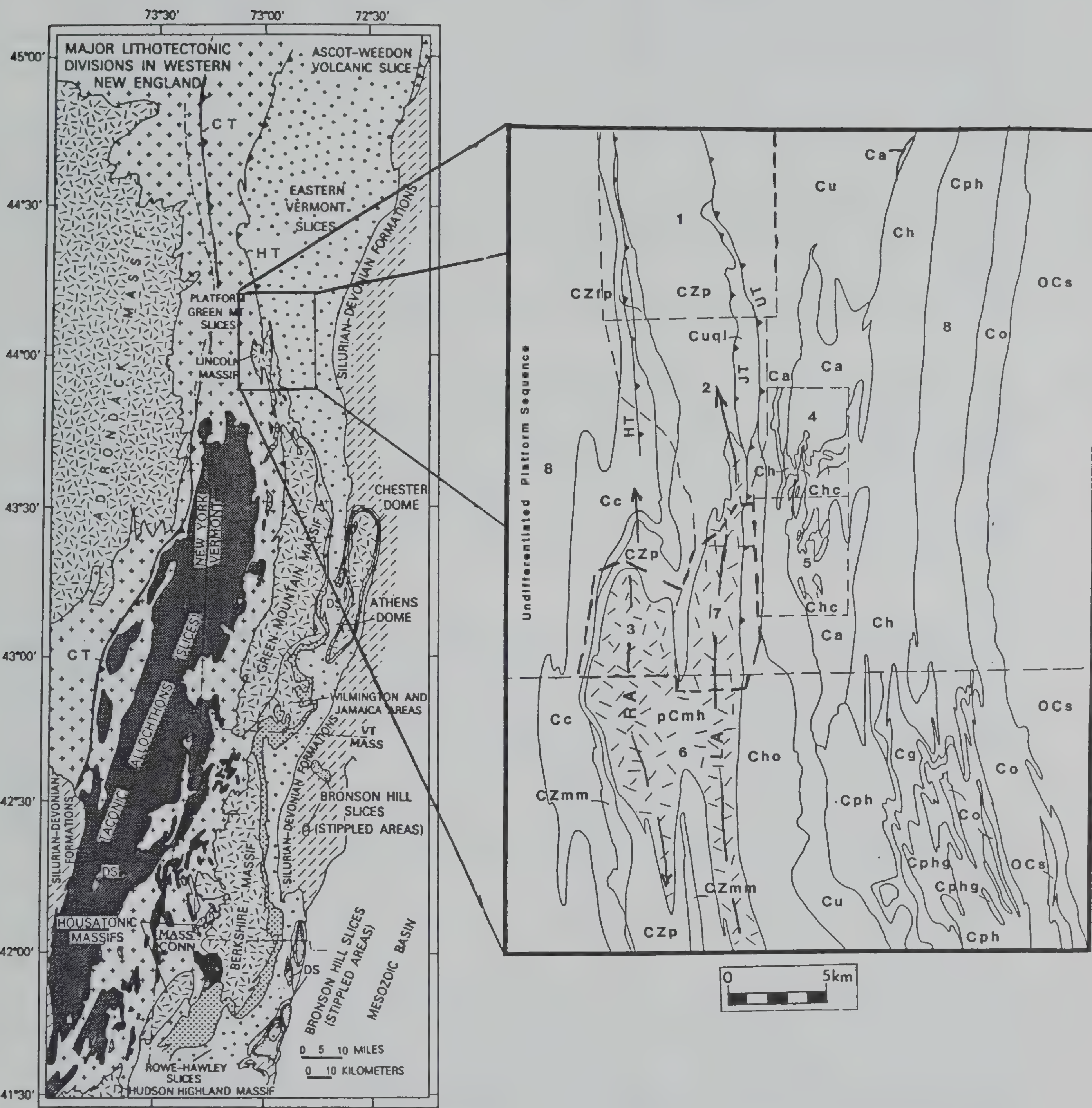
The study area is located in a topographic basin, directly west of the high peaks of the Green Mountains, between the towns of Lincoln and Ripton in central Vermont (Fig. 1). The area is centered on the northern half of the Lincoln massif which consists of rocks of the Mt. Holly Complex. The eastern part of the Lincoln massif (ELM) separates the eastern and western cover sequences across the axis of the Lincoln anticline. The Western Sequence mantles the basement rocks of the western part of the Lincoln massif (WLM) in the core of the Ripton anticline.

STRATIGRAPHY

Introduction

Middle Proterozoic rocks of the Mt. Holly Complex underwent a complicated depositional and tectonic history prior to Paleozoic mountain building. These rocks are exposed in the core of the Ripton and Lincoln anticlines and are separated by a major north-plunging syncline which contains Late Proterozoic/Lower Cambrian rocks of the Pinnacle Formation (Pl. 1). These two basement terranes are referred to as the western and eastern parts of the Lincoln massif (WLM and ELM). The Late Proterozoic to Lower Cambrian metaclastic rocks that mantle the Lincoln massif are separated into eastern and western sequences which are in fault contact across the axis of the Lincoln anticline. The Pinnacle, Fairfield Pond and Cheshire Formations of the Western Sequence

Figure 1. Generalized lithotectonic map of New England (Stanley and Ratcliffe, 1983; 1985) showing the distribution of basement massifs in northern New England (random dashes), the autochthonous platform sequence (plus signs) and the allochthonous eastern Vermont slices (open circles). The Taconic allochthons are shown in black. The Lincoln massif is the northernmost exposure of Middle Proterozoic basement in Vermont. Inset compilation map shows the location of the study area (bold outline), and the extent of detailed mapping in central Vermont prior to this study. Numbers refer to the following authors: 1-DiPietro (1983); 2-Tauvers (1982); 3-Prahl (1985); 4-O'Loughlin and Stanley (1986); 5-Lapp and Stanley (1986); 6-Osberg (1952); 7-DelloRusso (1986); 8-Doll and others (1961) and Cady and others (1963). Formation symbols are as follows: pCmh - Mount Holly Complex; CZp-Pinnacle Formation; CZmm - Moosalamoo Phyllite; CZfp - Fairfield Pond Formation; Cc - Cheshire Formation; Cho - Hoosac Formation; Cu - Underhill Formation; Cuql - quartz laminated schist of the Underhill Formation of Tauvers (1982) and DiPietro (1983); Ca - Mount Abraham Schist; Ch - Hazens Notch Formation; Chc - carbonaceous schist of the Hazens Notch Formation; Cg-Granville Formation; Cph - Pinney Hollow Formation; Cphg - greenstone in the Pinney Hollow Formation; Co - Ottauquechee Formation; OCs - Stowe Formation. UT represents the Underhill thrust zone, JT represents the Jerusalem thrust zone, RA represents the Ripton anticline, and LA represents the Lincoln anticline.



MAJOR LITHOTECTONIC DIVISIONS IN WESTERN NEW ENGLAND
ASCOT-WEEDON VOLCANIC SLICE
EASTERN VERMONT SLICES
PLATFORM GREEN MT SLICES
LINCOLN MASSIF
ADIRONDACK MASSIF
SILURIAN-DEVONIAN FORMATIONS
NEW YORK VERMONT
TACONIC ALLOCHTHONS (SLICES)
GREEN MOUNTAIN MASSIF
CHESTER DOME
ATHENS DOME
WILMINGTON AND JAMAICA AREAS
VT MASS
BRONSON HILL SLICES (STIPPLED AREAS)
HOUSATONIC MASSIFS
MASS CONN
BERKSHIRE MASSIF
MESOZOIC BASIN
ROWE-HAWLEY SLICES
HUDSON HIGHLAND MASSIF
0 5 10 MILES
0 10 KILOMETERS
Undifferentiated Platform Sequence
0 5km

comprise a continuous depositional section that records sedimentation during the rifting of the western margin of Iapetus and the development of a stable, passive continental margin. The Hoosac and Underhill Formations of the Eastern Sequence are believed to represent an eastern facies of the Pinnacle Formation.

Mt. Holly Complex

Twelve lithologic units are differentiated within the Mt. Holly Complex (Tauvers, 1982a; Prahl, 1985; DelloRusso, 1986a). Significant differences exist between the stratigraphy of the western and eastern parts of the Lincoln massif.

Western Part of the Lincoln Massif (WLM)

The Middle Proterozoic rocks of the WLM were mapped and described by Prahl (1985, Pl. 1, Fig. 2) who subdivided the Mt. Holly Complex into six lithologic units. The most widespread of these units is a granitic gneiss (Ymhg). Interlayered with the granitic gneiss are mappable units of grey quartzite (Ymhgq), biotite gneiss (Ymhbg), and a very distinctive microcline augen gneiss (Ymhga). Isolated exposures of amphibolite occur within the granitic gneiss and several exposures of massive, blue-grey quartzite (Ymhq) have been mapped along the western margin of this unit. These blue-grey quartzites are not continuous with the interlayered quartzite units within the granitic gneiss.

<u>Granitic gneiss (Ymhg):</u> This unit underlies most of the western part of the Lincoln massif. The lithology is a pink-grey, light-grey to greenish-white, rusty weathering, medium- to coarse-grained sericite-chlorite-biotite-microcline-quartz-plagioclase gneiss. Locally, the gneiss contains minor amounts of garnet and tourmaline. The gneiss displays compositional layering near its contact with the augen gneiss unit (Ymhga).

Map-scale layering is outlined in the gneiss by laterally continuous horizons of augen gneiss, biotite gneiss, and quartzite (Pl. 1). The presence of bedded quartzite within the granitic gneiss indicates that the gneiss most likely has a sedimentary origin.

<u>Microcline augen gneiss (Ymhga):</u> This unit is a very distinctive, light- to medium-grey to white, coarse-grained, biotite-quartz-plagioclase-microcline augen gneiss, containing minor amounts of chlorite, sericite, epidote, tourmaline, sphene and opaques (Prahl, 1985). The rock contains abundant microcline augen up to 4 cm long. It is well-exposed at numerous localities which define a thin unit within the granitic gneiss (Ymhg). As a result of the lateral continuity of this unit and its distinctive lithology, several map-scale, Paleozoic folds have been defined within this part of the massif.

The contacts between the augen gneiss and the surrounding granitic gneiss appear gradational over a distance of 0.5 to 2 m (Prahl, 1985). The parallelism in outcrop pattern between the augen gneiss and the bedded

quartzite (Ymhgq) suggests that they are part of a continuous pre-Grenvillian sedimentary sequence.

__Quartzite (Ymhgq):__ This unit is exposed in a narrow but laterally continuous horizon within the granitic gneiss (Ymhg). It is a grey to white, massively bedded quartzite containing minor amounts of sericite, chlorite, magnetite, biotite, and tourmaline (Prahl, 1985). Beds range from 30 cm to 5 m in thickness.

Contacts with the surrounding granitic gneiss (Ymhg) and biotite gneiss (Ymhbg) are gradational. Beds of quartzite are also found within the biotite gneiss (Ymhbg). The composition and bedded nature of the quartzite clearly defines its sedimentary origin. More importantly, the conformable association of the quartzite with surrounding gneisses suggests a similar origin for those rocks as well.

__Biotite gneiss (Ymhbg):__ This unit is exposed in several outcrops in the western part of the WLM and is laterally discontinuous. The lithology is a dark-grey, rusty weathering, coarse-grained quartz-plagioclase-microcline-biotite gneiss containing minor amounts of chlorite and sericite (Prahl, 1985). Magnetite porphyroblasts up to 1 cm in size are common locally. Segregations of biotite give the rock a streaked appearance. Significantly, this unit contains continuous layers of quartzite (Ymhgq) which indicate a sedimentary origin. Contacts with the surrounding gneiss are gradational.

__Amphibolite (Ymha):__ This unit is exposed in several isolated outcrops within the granitic gneiss. The rock is a dark bluish-black to greenish-black, fine- to medium-grained, massive hornblende-plagioclase amphibolite. It is commonly altered to plagioclase-chlorite-epidote-biotite schist. Pyroxene has been reported in this lithology by Prahl (1985). These amphibolites are lithologically similar to those described by Tauvers (1982a, 1982b) in the South Lincoln area. Due to the isolated and massive occurrence of this lithology, the nature of the origin of the amphibolite and its relation to surrounding units remains unknown.

__Blue quartzite (Ymhq):__ Three large bodies of this unit are exposed along the western margin of the WLM (Pl. 1). The rock is a clean, massive, light-blue-grey quartzite containing minor amounts of mica and opaques (Prahl, 1985). The unit typically displays beds 0.5 to 1 m thick.

The contacts between the blue quartzite and granitic gneiss are unexposed, but have been interpreted as gradational by Prahl (1985). A sharp contact between the blue quartzite (Ymhq) and granitic gneiss has been documented by DelloRusso (1986a) along the western margin of the ELM. Because these quartzite bodies are located along the margins of the granitic gneiss, it is possible that they were deposited upon the gneiss rather than as part of a depositional sequence that includes the gneiss. This would imply that the blue quartzite/gneiss contact is an unconformity. The quartzite is unconformably overlain by cobble conglomerate of the basal Pinnacle Formation which contains quartzite cobbles.

Eastern Part of the Lincoln Massif (ELM)

The rocks of the Mt. Holly Complex within the ELM are divided into three distinct lithic assemblages. The most widespread unit is granitic gneiss (Ymhg, Pl. 1, Pl. 2). The gneiss is also associated with numerous biotite schist zones (Ymhgb) and intruded by post-Grenvillian, metamorphosed, mafic dikes (Zmd). Two metasedimentary sequences are mapped overlying the granitic gneiss. A sequence of blue quartzite (Ymhq), tourmaline-chloritoid schist (Ymht), and layered amphibolite (Ymhml) is exposed along the western margin of the ELM. Along the eastern margin of the massif, a sequence of plagioclase augen gneiss (Ymhag) and schistose quartzite (Ymhaq) is recognized.

Granitic gneiss (Ymhg): This unit comprises the majority of the ELM. The rock is a light-grey to white, fine- to medium-grained muscovite-perthite-microcline-quartz-plagioclase gneiss. The gneiss is typically fine to medium-grained and weathers light-grey to white. Biotite is rare within this unit. Garnet, chlorite, magnetite and zircon are also rarely observed. The gneiss is at least partially altered everywhere and contains varying amounts of sericite, epidote, and calcite which are a result of Paleozoic retrogression of feldspar. Similarly, Tauvers (1982a) described this unit in the northernmost part of the ELM as a grey, greenish-white, fine- to medium-grained sericite-microcline-quartz-plagioclase gneiss and emphasized the relative plagioclase-rich composition of the gneiss in that area. Comparisons of modal analyses of gneisses from the western and southern parts of the Lincoln massif (Prahl, 1985; Osberg, 1952) are more microcline- and biotite-rich than the granitic gneiss (Ymhg) of the ELM (DelloRusso, 1986a).

Compositional layering is typically not observed in the gneiss. Rare banding is observed in the southwestern part of the ELM where the gneiss is more microcline-rich (Pl. 2, Loc. EE8). The gneiss in this area may be similar to the microcline-rich gneisses described by Osberg (1952) in the Ripton area and the layered gneisses described by Prahl (1985) in the WLM. The granitic gneiss throughout the rest of the ELM is relatively homogeneous and typically rich in plagioclase (30-50%; Tauvers, 1982a; DelloRusso, 1986a).

The unconformity between the gneiss and the overlying Pinnacle Formation to the west is well-exposed at several localities (Pl. 2, Locs. P9, T10, S10, FF8). Locally, the unconformity is masked by the similar appearance of the basal Pinnacle Formation and the gneiss. This similarity is a result of compositional and subsequent Paleozoic metamorphism. The contact can be distinguished, however, by the presence of rounded blue quartz clasts and rare gneiss boulders.

The unconformity along the eastern margin of the gneiss is not exposed, but is constrained by closely-spaced outcrops of Mt. Holly gneiss and the muscovite-quartz-feldspar metagreywacke (CZhg) of the overlying Hoosac Formation. Contacts between the gneiss and all other lithic units in the Mt. Holly Complex, where exposed, are sharp.

The granitic gneiss (Ymhg) is interpreted as an orthogneiss, because its composition is relatively homogeneous and it lacks any obvious compositional layering. Although presently mapped as the same unit, the gneiss in the

southwestern part of the study area may be of sedimentary origin because it contains compositional layering which is similar to the gneisses of the WLM. Although metasedimentary units are described in the Mt. Holly Complex of the ELM, these units do not occur within the granitic gneiss (Ymhg) and, therefore, do not necessitate a similar origin for the gneiss. An igneous origin for the granitic gneiss of the ELM was also proposed by Tauvers (1982a, 1982b) because it is locally rich in plagioclase.

Amphibolite (Ymha): Several amphibolite bodies (Ymha) occur within the gneiss (Ymhg) of the northern part of the ELM (Pl. 1). The rock is a dark green, fine- to medium-grained, massive, hornblende-plagioclase amphibolite. It is commonly altered to a plagioclase-chlorite-epidote-biotite-amphibole schist (Tauvers, 1982a). Coexisting blue-green amphibole and actinolite have been described by Tauvers (1982b) in this unit. The amphibolite occurs within, and in sharp contact with, the granitic gneiss (Ymhg) and is interpreted to be intrusive. On the basis of composition and their close association with the granitic gneiss, these amphibolites are correlated with those mapped in the western part of the massif.

Biotite schist (Ymhgb): This unit is a fine-grained, quartz-sericite-biotite schist containing minor amounts of albite, chlorite, epidote, carbonate and opaques. Medium-grained, randomly-oriented biotite porphyroblasts are abundant in this lithology. Narrow zones of strongly foliated biotite schist occur within, and in contact with, the granitic gneiss (Pl. 2).

Three hypotheses have been proposed for the origin of the biotite-rich schists (DelloRusso, 1986a): 1) biotite enrichment of the granitic gneiss associated with the infiltration of potassium-, iron-, and magnesium-rich fluids along shear zones, 2) retrogression of amphibolite horizons in the gneiss (Ymhg), and 3) retrogression or metasomatization of mafic dikes (Zmd).

Plagioclase augen gneiss (Ymhag)/schistose quartzite (Ymhaq) sequence: The augen gneiss is a coarse-grained sericite-chlorite-quartz-biotite-plagioclase augen gneiss. Large white plagioclase porphyroclasts up to 4 cm in diameter are observed at location X16 (Pl. 2). The abundant dark matrix which surrounds these clasts is biotite-rich and contains small clasts of bluish-grey to grey quartz. Chlorite, epidote, and sericite are also common in the matrix.

Fine-grained augen gneiss is interbedded with schistose quartzite and conglomeratic quartzite in the eastern part of the sequence (Pl. 2, Loc. X17-18). The quartzite has a blue-green color and locally contains up to 22% chlorite and 25% sericite. Fine-grained plagioclase-rich augen gneiss is exposed east of the schistose quartzite. This gneiss comprises the easternmost exposures of the Mt. Holly Complex in the ELM, with the exception of fault slivers in the Hoosac Formation. The fine-grained augen gneiss weathers white to light-grey with porphyroclasts of plagioclase less than 1 cm in diameter. The matrix is largely sericite, epidote, and quartz with less than 6% biotite. Rare clasts of quartzo-feldspathic gneiss, similar to the granitic gneiss (Ymhg), are observed within this lithology indicating that the

granitic gneiss is older and that it is the likely source for the abundant quartz and feldspar in the augen gneiss (DelloRusso, 1986a).

The abundance of biotite in the matrix of the coarse-grained augen gneiss and the abundance of chlorite in the schistose quartzite indicates a mafic source for these two units. The large body of amphibolite (Ymhml) exposed just west of the sequence may represent a remnant of such a source.

The contact between the granitic gneiss (Ymhg) and the plagioclase augen gneiss (Ymhag) is not exposed. The basement-cover unconformity between the augen gneiss and the Hoosac Formation to the east is not exposed, but can be constrained by closely-spaced exposures of fine-grained feldspar augen gneiss and biotite-rich quartz-feldspar metawacke (CZhbg) of the Hoosac Formation.

The presence of large feldspar clasts, quartz clasts, rare gneiss clasts, and interbedded schistose quartzite clearly indicates a sedimentary origin for this sequence. These relations suggest that an unconformity exists between the granitic gneiss (Ymhg) and the plagioclase augen gneiss (Ymhag).

<u>Blue quartzite (Ymhq)/tourmaline-chloritoid schist (Ymht) sequence:</u> This sequence is well-exposed as a large fold along the northwestern margin of the ELM. The quartzite is a clean, massive, light blue to light grey in color and contains less than 10% sericite. The quartzite displays beds 0.5 to 1 m thick which are warped into broad folds. Bedding planes in massive exposures are defined by dark inclusion traces of micas and opaques. Mullions are well developed on bedding surfaces and are parallel to the hinges of the broad, open folds within the quartzite.

The quartzite is interbedded with a quartz-chloritoid-tourmaline-sericite schist (Ymht). Abundant fine-grained needles of tourmaline and large porphyroblasts of chloritoid up to 1 cm in diameter are characteristic of this lithology. This unit typically contains greater than 60% sericite. Lamellar-twinned aggregates of chloritoid form radial patterns which were described as "suns" or "rosettes" by Brace (1953). Graphite is common in minor amounts and occurs rarely as pure, 3-5 cm thick seams parallel to bedding. Massive blue quartzite is also observed along the western basement-cover contact at locations U9 and GG5 (Pl. 2).

It is uncertain whether the blue quartzite described above is correlative to the chloritic quartzite (Ymhaq) within the plagioclase augen gneiss (Ymhag) along the eastern margin of the massif. It is possible that the two units had a similar quartz-rich source, but clearly the blue quartzite (Ymhq) did not share the mafic source component of the augen gneiss (Ymhag).

Tourmaline-chloritoid-sericite schist is restricted to the large quartzite body in the northwest part of the study area. The schist is a continuous map unit and is useful in deciphering the folded pattern in the quartzite body. The iron-alumina-rich composition suggests that the rock may have originated as a lateritic soil. Deposits such as these form by weathering in wet tropical climates where aluminum and iron are concentrated in the soil (Reading, 1978). The occurrence of graphite seams suggests that organic material was present in the protolith.

The fact that these quartzite bodies and their associated paleosols are located along the margins of the massif suggests that they were deposited on top of the gneiss. The contact between the blue quartzite (Ymhq) and the granitic orthogneiss is sharp and only exposed at location M13 (Pl. 2). This contact, therefore, is interpreted to be an unconformity. The unconformity between the quartzite-schist-amphibolite sequence and the lower Pinnacle Formation to the west is not exposed, but is controlled by outcrops spaced greater than 30 m apart.

<u>Layered mafic schist (Ymhml):</u> Large exposures of this unit are present at locations K11-12 and L11-12. The rock is a dark-green to black, medium- to coarse-grained, garnet-plagioclase-hornblende amphibolite which is largely retrograded to epidote-chlorite-quartz-biotite-actinolite-plagioclase schist as a result of Paleozoic metamorphism. Large euhedral magnetite porphyroblasts are locally common in the retrograded amphibolite. Sericite, calcite and sphene are rare.

At location L12 (Pl. 2) the rock is fine- to medium-grained and displays a sigmoidal compositional layering. The compositional layering is defined by the segregation of hornblende and plagioclase. Garnet is scattered throughout and occurs in equilibrium with the plagioclase and hornblende (DelloRusso, 1986a).

A second body of amphibolite (Ymhml) is found at location X15-Y15 (Pl. 2). This body is smaller than the body described above and occurs along the contact between granitic gneiss and plagioclase augen gneiss. It is very coarse-grained and displays minor alteration of the primary textures in the rock. Rare chlorite is observed along some amphibole grain boundaries. Plagioclase displays albite twinning with only minor alteration to epidote.

Amphibolite units (Ymha, Pl. 1) described by Tauvers (1982a) and Prahl (1985) occur almost exclusively within the granitic gneiss (Ymhg). In contrast, the layered mafic schist (Ymhl) is present along the periphery of the massif. This suggests that at least two distinct populations of mafic rocks occur within the massif.

The contacts between amphibolite and the surrounding granitic gneiss and quartzite are not exposed. Xenoliths of granitic gneiss (Ymhg) are observed, however, within the layered mafic schist (Ymhml) at location L12 (Pl. 2). The xenoliths indicate that the mafic schist (Ymhml) is younger than the gneiss. The location of the layered mafic schist along the periphery of the massif suggests that the mafic schist may have been extruded on to the gneiss (Ymhg). At location L12 (Pl. 2), a granitic pegmatite cuts the layered mafic schist (Ymhml). The fact that these pegmatites have not been found in either the quartzite (Ymhq) or the tourmaline-chloritoid schist (Ymht) suggests that these units were deposited on the layered mafic schist.

<u>Mafic dikes (Zmd):</u> These dikes occur as quartz-sericite-epidote-biotite-chlorite-plagioclase schist with minor amounts of carbonate, sphene and opaques. These minerals represent a Paleozoic retrogressive assemblage. The dike rocks are dark-green-brown weathering, fine- to medium-grained, and do

not display a strong foliation except where they are associated with faults. No primary amphibole is observed. Where carbonate has been weathered, the rock is marked by rusty-specks.

Euhedral plagioclase phenocrysts up to 2 cm in length are characteristic of these intrusives and represent relicts of the primary igneous texture of the rock. These phenocrysts are locally abundant and comprise up to 75% of the rock (Pl. 2, Locs. P10, R15).

Wide zones of biotite-rich schist commonly occur along the dike margins (Pl. 2, Loc. Q14) suggesting that the dikes were metasomatically altered along the contact zone. These zones are analogous to the many, narrow, biotite schist zones (Ymhgb) previously described.

At locations P10, Q10, and M13 (Pl. 2), dike contacts with the granitic gneiss (Ymhg) and blue quartzite (Ymhq) are well exposed. They are sharp and truncate the Grenvillian fabric observed in the adjacent gneiss. At location Q10 a partially detached appendage of gneiss is enveloped in the dike rock. Xenoliths of gneiss (Ymhg) are also observed at locations S10 and P10 (Pl. 2). At location M13 (Pl. 2), a sliver of quartzite is partially detached along the sharp dike contact. These relationships document an intrusive origin for the dikes. These dikes are deformed and truncated by Paleozoic fault zones. Detailed mapping indicates that the dikes are truncated by the basement-cover contact along the western margin of the massif (Pl. 2, Loc. P9-10).

The observed relationships suggest that these mafic dikes represent Late Proterozoic rift-stage intrusives. Considering the large time span between the Grenvillian orogeny and the rifting of Iapetus (approx. 400 m.y.), other post-Grenvillian events cannot be excluded.

The mafic dikes documented in this study may represent feeder dikes related to the Tibbit Hill Volcanic Member of the Pinnacle Formation. Recent geochemical analyses of the Tibbit Hill rocks of northern Vermont indicate that they are rift volcanics (Coish and others, 1985). Geochemical data from the mafic dikes of this study (Filosof, 1986) reveal that they are high-titanium, high-phosphorus, alkaline basalts that correlate well in composition with the Tibbit Hill volcanic rocks.

Western Sequence

The Western Sequence is represented by the Pinnacle, Fairfield Pond and Cheshire Formations (Pl. 1, Fig. 2). These formations record the depositional history of the western margin of Iapetus from the rift stage to the development of a stable continental shelf during Late Proterozoic to Early Cambrian time. The Cheshire and Fairfield Pond Formations are not subdivided on the geologic maps (Pl. 1). Five distinct lithofacies are recognized within the Pinnacle Formation.

Pinnacle Formation

The rocks of the Pinnacle Formation represent immature rift-clastic sediments that were shed from local topographic highs of the Mt. Holly Complex

during Late Proterozoic rifting. The Pinnacle Formation is differentiated into five lithic units (Pl. 1): conglomerate (CZpbc), muscovite metawacke (CZpm), biotite metawacke (CZpbg), Forestdale dolarenite (CZpfd), and chlorite-magnetite schist (CZpcl). Each of these lithologies was described in detail by Tauvers (1982a, 1982b) in the Lincoln area directly north of the Lincoln massif. Prahl (1985) recognized the five lithofacies along the western limb of the Ripton anticline. DelloRusso (1986a) also described the lowermost units of the Pinnacle Formation (CZpbc and CZpm) exposed along the western limb of the Lincoln anticline.

<u>Conglomerate (CZpbc):</u> Coarse cobble and boulder conglomerate represents the basal unit of the Pinnacle Formation which is well exposed at numerous localities along the western margins of the ELM and WLM. Conglomerate has also been documented in stratigraphically higher parts of the Pinnacle Formation by Prahl (1985) and DiPietro (1983a, 1983b). The conglomerate commonly occurs in direct contact with the basement gneiss, thus defining the basement-cover unconformity. The lithology is a poorly sorted, matrix-supported, quartz and gneiss cobble conglomerate, with cobbles up to 50 cm in diameter. Rare quartzite boulders up to 3 m in length have been observed by Prahl (1985). The matrix weathers grey-brown to light-grey and is composed of feldspar-quartz-biotite-muscovite-chlorite schist containing minor amounts of calcite and magnetite. Rare boulders of gneiss up to 75 cm in diameter are exposed at locations O10 and T10 (Pl. 2) and lie directly on the unconformity. Bedding is discernable by rare, thin, coarse-grained layers surrounded by the finer-grained, schistose matrix. The unconformable contact between the basal conglomerate and the granitic gneiss (Ymhg) is well exposed in several outcrops as a sharp depositional contact with no evidence of faulting.

The basal cobble and boulder conglomerates are laterally discontinuous and probably represent the high energy deposits of the proximal reaches of subaerial and subaqueous fans which filled a rift basin during Late Proterozoic time (Tauvers, 1982a, 1982b).

<u>Muscovite metawacke (CZpm):</u> This unit overlies the basal conglomerate (CZpbc) along the western limb of the Lincoln anticline. At the northern end of the ELM, Tauvers (1982) described this lithology as a mottled muscovite schist. He recognized a similar lithology overlying the biotite metawacke member of the Pinnacle Formation which he referred to as muscovite greywacke.

The metawacke is silvery, light-grey to grey-green, rusty-brown weathering, fine-grained, massive quartz-feldspar-muscovite metawacke. It contains minor amounts of chlorite, magnetite, epidote and tourmaline. Biotite is uncommon and rock fragments are conspicuously absent. Locally, this unit displays well-developed 5 mm thick quartz-feldspar laminations. Bedding is locally defined in this unit by isolated layers of rounded, blue quartz pebbles. Relict detrital grains of blue quartz and feldspar are common and are an important criteria used in distinguishing this lithology from the adjacent granitic gneiss (Ymhg) of the basement. Detrital feldspar grains, including microcline and perthite, within the metawacke clearly indicate that the granitic gneiss (Ymhg) was the source of detritus for this unit. The upper contact of this unit is gradational at the northern end of the study area (Tauvers, 1982a, 1982b) and is defined by an increase in the percentage of

biotite in the metawacke, as well as by an increase in grain size. The contact with the underlying conglomerate is also gradational because the matrix of the conglomerate is identical in composition and appearance to the muscovite metawacke. The biotite-poor lithology is only locally recognized by Prahl (1985) in the western part of the area where it is the dominant lithofacies.

The muscovite metawacke may represent a finer-grained, more distal facies than the conglomerate, which was probably deposited during periods of relative quiescence during the rift stage when basement relief was less.

Biotite metawacke (CZpbg): This unit is the most widespread lithology of the Pinnacle Formation. The biotite metawacke was described north of the ELM by Tauvers (1982a, 1982b) and mapped around the WLM by Prahl (1985). Biotite metawacke is not exposed along the western limb of the Lincoln anticline although it is interpreted to overlie the muscovite metawacke (CZpm) in that area.

The lithology is a dark-grey, carbonate-mottled, brown-weathering, medium-to coarse grained, massively bedded quartz-feldspar-muscovite-chlorite-biotite metawacke. Locally this unit contains conglomeratic beds up to 1 m thick with well rounded quartz and gneiss cobbles up to 5 cm in diameter (Tauvers, 1982a, 1982b; Prahl, 1985). Contacts with surrounding lithologies are generally gradational. They are defined by an increase in chlorite and magnetite and by the lack of biotite in the overlying chlorite-magnetite schist (CZpcl) and the lack of biotite and conglomerate in the underlying muscovite schist (CZpm).

The presence of conglomerates and the relative coarse-grained nature of the biotite metawacke suggests a period of renewed tectonic activity within the rift, in contrast to the overlying and underlying muscovite metawacke.

Chlorite-magnetite schist (CZpcl): This unit was described by Tauvers (1982a, 1982b) north of the Lincoln massif. He designated this unit as the "Upper Pinnacle" in the Lincoln area. A similar lithofacies has been described by Prahl (1985) in a stratigraphically lower part of the Pinnacle Formation along the western limb of the Ripton anticline (Pl. 1). The lithology is a light-green, locally rusty, silvery-grey weathering, fine-grained quartz-plagioclase-muscovite-magnetite-chlorite schist. Abundant euhedral magnetite porphyroblasts up to 7 mm in diameter are diagnostic of this unit. Locally, the schist is interbedded with chlorite-rich, pebbly metawacke. Discontinuous pods of vein quartz and interbeds of dolostone are common. The contacts with the overlying grey phyllite of the Fairfield Pond Formation and the underlying biotite and muscovite metawackes of the lower Pinnacle are gradational and locally difficult to delineate. Along the western limb of the Ripton anticline, the chlorite-magnetite schist lies unconformably upon the basement gneiss (Ymhg) and is overlain by muscovite metawacke (Prahl, 1985).

Forestdale dolarenite (CZpfd): This unit occurs entirely within the chlorite-magnetite schist member (CZpcl) of the Pinnacle Formation. North of the Lincoln massif, the dolarenite occurs in the upper part of the Pinnacle Formation (Tauvers, 1982a, 1982b). Prahl (1985) has also documented

dolarenite within the lower part of this formation, along the western margin of the WLM. The dolarenite is a light-brown, dark-brown to black weathering, sandy dolarenite (Tauvers, 1982a, 1982b; Prahl, 1985). It is typically massively bedded containing thin chlorite schist interbeds. Subangular clasts of quartz and feldspar up to 7 mm in diameter are common. Beds up to 0.5 m thick exhibit normal grading of clastic grains (Tauvers, 1982a, 1982b). Outcrops of the dolarenite commonly form ridges up to 23 m high (Prahl, 1985). A "transitional" facies of the dolarenite is described by Tauvers (1982a, 1982b) in the Lincoln area. This facies consists of a thinly-bedded, sandy dolarenite intercalated with up to 50% chlorite schist. These intercalations illustrate the gradational nature of the upper and lower contacts between the dolarenite and surrounding chlorite-magnetite schist (CZpcl).

The carbonate material in the dolarenite has been interpreted to have been derived from localized carbonate banks that formed on basement horst blocks during the later stage of rifting (Tauvers, 1982a, 1982b). The presence of poorly-sorted, angular detrital quartz and feldspar grains within the massive dolarenite, lithic fragments of intraformational chert within the "transitional facies" of the dolarenite, and polymictic pebble conglomerate in the overlying chlorite-magnetite schist indicates a resedimented origin for the Forestdale dolarenite (Tauvers, 1982a, 1982b).

Fairfield Pond Formation

<u>**Chlorite phyllite (CZFp):**</u> The lithology is a grey to grey-green, light-grey to light-brown weathering, fine-grained quartz-sericite-chlorite-phyllite. Biotite is characteristically rare. The rock displays fine, light-grey quartz-rich laminations that are typical of this lithology. Magnetite is common near the base of the unit. Alternating quartz/sericite-chlorite laminations are up to 2.5 cm thick and are interpreted as primary, quartz-rich silt beds interlaminated with shale. The Fairfield Pond phyllite grades upward into the lower argillaceous quartzite of the Cheshire Formation and is thought to represent a transitional (rift-drift) shallowing-upward sequence between the Pinnacle and Cheshire Formations (Tauvers, 1982a, 1982b).

The Fairfield Pond Formation was originally considered to be a member of the Underhill Formation (Doll and others, 1961). The Fairfield Pond Formation is correlative to the West Sutton Formation in northern Vermont (Booth, 1950), and to the Moosalamoo Member of the Mendon Formation in the East Middlebury-Rutland area (Osberg, 1952; Brace, 1953). Recent work by Tauvers (1982a, 1982b) and DiPietro (1983a, 1983b) have shown that the stratigraphy across the Lincoln anticline is not symmetrical in the Lincoln area and that the Underhill/Pinnacle contact of the Eastern Sequence is a major thrust fault. As a result, Tauvers (1982a, 1982b) interpreted the Underhill Formation as an eastern facies of the Pinnacle Formation and suggested that the Fairfield Pond Member of the Underhill Formation (Doll and others, 1961) be raised to formation status. The nomenclature of Tauvers (1982a, 1982b) is followed in this report.

The grey phyllite of the Fairfield Pond Formation is interpreted as a fine-grained, transitional facies that separates the rift clastics rocks of

the Pinnacle Formation and the drift-facies quartzites of the Cheshire Formation (Tauvers, 1982a, 1982b).

Cheshire Formation

Massive quartzite (Cc): The Cheshire Formation, which contains the trilobite Olenellus, is Lower Cambrian in age and is the youngest unit in the study area (Walcott, 1888). The Cheshire Formation was designated as the western limit of the study area and thus only the lower part of the formation is described and represented on Plate 1.

The lithology is a white, grey to pink, massive, fine- to medium-grained quartzite and argillaceous quartzite containing some sericite phyllite (Tauvers, 1982a, 1982b; Prahl, 1985). Discontinuous, mottled, white, rippled quartzite beds from 2.5 to 20 cm thick occur locally (Tauvers, 1982a, 1982b).

The Cheshire Formation is interpreted to be of shallow water marine origin (Myrow, 1982; Tauvers, 1982a, 1982b), representing the development of a stable, passive continental margin and the end of the rift stage during Early Cambrian time.

Eastern Sequence

The Eastern Sequence is represented by the Hoosac and Underhill Formations along the eastern margin of the Lincoln massif. These rocks represent an eastern facies of the Western Sequence rift-clastic rocks. The rocks of the Underhill Formation have similarities to the rocks of the Hoosac Formation and are interpreted as an eastern facies equivalent of the Hoosac Formation (Tauvers, 1982a, 1982b).

Hoosac Formation

The Hoosac Formation is subdivided into four units: schistose metawacke (CZhg), biotite metawacke (CZhbg), conglomerate (CZhbc), and mafic schist (CZhms). All units lie to the west of the undifferentiated schists of the Underhill Formation.

Biotite metawacke (CZhbg) and conglomerate (CZhbc): The biotite metawacke (CZhbg) is a fine- to medium-grained, biotite-rich, chlorite-quartz-plagioclase schistose metawacke. Biotite schist occurs locally. Chlorite is common and locally abundant. Sericite, epidote, carbonate and opaques are present in minor amounts. The biotite metawacke is identical in composition to the matrix of the conglomerate (CZhbc) but is differentiated on Plate 2 by the lack of coarse clastic material in the biotite metawacke.

The conglomerate (CZhbc) is interpreted as a local coarse-grained deposit in the biotite metawacke (CZhbg). The conglomerate is observed at the South Lincoln bridge outcrop (Pl. 2, Loc. D19) and at isolated outcrops in the southeastern part of the study area (Pl. 2, Loc. X20). The conglomerate at South Lincoln is exposed in the river below the bridge and along the eastern bank just south of the bridge. This conglomerate was originally mapped by Tauvers (1982a, 1982b). It is in depositional contact with a sliver of

quartz-rich Mt. Holly gneiss. The gneiss displays a mylonitic fabric and cobbles in the conglomerate are stretched up to 30 cm in length (Strehle and Stanley, 1986). Most cobbles at this locality are quartzite although gneiss cobbles are also present (Tauvers, 1982a, 1982b).

South of South Lincoln bridge, conglomerate is rarely observed and, where present, appears thin and discontinuous. The conglomerate at location X20 (Pl. 2) contains primarily felsic clasts which do not exceed 2 cm in diameter. Southern exposures of the conglomerate also appears to be more biotite-rich than at South Lincoln.

Contacts of the conglomerate and metawacke (CZhbg) with surrounding units are not exposed and therefore the contact relationships are uncertain. Locally, some contacts are thought to be faulted and others are interpreted to be depositional. The biotite metawacke (CZhbg) and conglomerate (CZhbc) are interpreted to be a facies equivalent of the schistose metawacke (CZhg). No gradation, however, has been documented in the field due to sparse outcrop.

The biotite metawacke appears to be a laterally discontinuous unit which may reflect changes along strike in source rock composition. The biotite-rich matrix of this lithology is interpreted to reflect a mafic source rock. Mafic rocks in the Mt. Holly Complex are the likely source for these biotite-rich metawackes and schists. The limited occurrence of the mafic rocks in the ELM may be reflected in the local occurrence of the biotite metawacke (CZhbg) along the eastern margin of the massif.

Mafic schist (CZhms): This unit is a biotite-chlorite-quartz-plagioclase-hornblende amphibolite which locally contains garnet, epidote, sphene and opaques. The amphibole is hornblende and occurs as fine-grained, blue-green (Z direction) needles aligned parallel to the dominant schistosity (S2) in the rock and display turbid cores. At location L21 (Pl. 2), a second population of pale to colorless amphibole coexists with the blue-green hornblende. Where hornblende is absent, the rock is rich in chlorite and biotite. Locally, quartz and plagioclase are strongly segregated from amphibole and mica, giving the rock a pin-striped appearance. Albite porphyroblasts commonly display an helicitic texture and appear to have been partially rolled in the dominant schistosity (pre- to syn-S2).

The mafic schist occurs as a continuous horizon within the schistose metawacke of the Hoosac Formation (DelloRusso, 1986a, 1986b). Exposures of mafic schist (CZhms) occur at locations X20, Q21-R21, and L21 (Pl. 2). The mafic schists at these localities are similar in overall appearance and are interlayered with quartzo-feldspathic metawackes (CZhg). This interlayering is visible both on the scale of the mapping and in hand specimen. The contacts between felsic and mafic layers are typically sharp and are interpreted to be depositional. The interlayering is believed to reflect a primary segregation in the rock. Considering the severity of Paleozoic deformation and metamorphism observed in these rocks, however, it is likely that bedding has been severely transposed.

A similar mafic schist, exposed at location B21 (Pl. 2), contains coarse-grained amphibolite and occurs along strike with the previously defined belt

of mafic rocks. It also occurs in a sequence of quartzo-feldspathic metawacke which suggests that the sequence may be part of the same stratigraphic horizon.

The mafic schist may be equivalent with the Tibbit Hill Volcanic Member of the Pinnacle Formation as well as with other greenstone bodies mapped within the Underhill Formation in northern Vermont (Doll and others, 1961; Coish and others, 1985). These mafic rocks may also be genetically related to the mafic dikes (Zmd) described within the ELM. The association of this unit with quartzo-feldspathic metawacke (CZhg) suggests that these rocks have an extrusive or volcanoclastic origin rather than an intrusive origin.

Schistose metawacke (CZhg): This lithology is a light-grey to white, locally rusty weathering, fine- to medium-grained, sericite-quartz-plagioclase metawacke containing minor amounts of garnet, biotite, chlorite and epidote. Garnet porphyroblasts are commonly chloritized. Biotite is rare in this unit. Locally the unit is very albitic and, in one exposure (Pl. 2, Loc. X20), albite porphyroblasts comprise approximately 80% of the rock.

The schistose metawacke (CZhg) is distinguished from the biotite metawacke (CZhbg) and conglomerate (CZhbc) by its lighter color, lack of biotite, and the absence of coarse clastic material. Conglomerates are not observed in the schistose metawacke (CZhg). This unit becomes more mica-rich eastward, toward the gradational contact with the schists of the Underhill Formation. The Underhill schists (CZu) are similar to the Hoosac rocks in that they are locally quartz-feldspar rich although overall the Underhill Formation is distinctly more white mica-rich than the Hoosac lithologies.

The contact between the Underhill Formation and the Hoosac metawacke (CZhg) is not well-defined and appears to be gradational over several hundred meters. Strongly foliated muscovite schists, typical of the Underhill Formation, are intercalated with the quartzo-feldspathic Hoosac lithologies along the contact zone. This contact therefore separates rocks that are dominantly quartz-feldspar metawackes from those that are muscovite-rich schists. The contacts between the schistose metawacke (CZhg) and the biotite metawacke (CZhbg)/conglomerate (CZhbc) units are not exposed.

Underhill Formation

The Underhill Formation is made up of a variety of lithologies that have complex stratigraphic and structural relationships (Tauvers, 1982a, 1982b; Lapp, 1985; Lapp and Stanley, 1986; O'Loughlin, 1986; O'Loughlin and Stanley, 1986). The lack of outcrop in the valley to the east of the study area does not allow a thorough investigation of the lithologic relations in this formation.

A distinctive lithology in the Underhill Formation, the quartz-laminated schist (CZuql; Tauvers, 1982a, 1982b), is recognized as a sliver along the Underhill thrust zone, and will be described separately below.

Muscovite schist (CZu): The composition of the muscovite schist ranges from muscovite-rich schist to quartzo-feldspathic metawacke. The most common

lithology in the Underhill Formation exposed in the study area is a silvery, rusty-weathering, medium-grained, garnetiferous chlorite-albite-quartz-muscovite schist containing minor amounts of biotite and epidote. Garnet porphyroblasts occur locally and are commonly chloritized.

Abundant, thin, discontinuous quartz pods are characteristic of the Underhill Formation in this area. These pods are typically folded or stretched parallel to the dominant schistosity. The rocks of the Underhill Formation are generally poorer in biotite and albite and are more quartz-rich than the Hoosac lithologies. Some quartz-feldspar rich facies in the Underhill Formation are indistinguishable from the Hoosac metawackes.

In the Starksboro and Lincoln townships, Tauvers (1982a, 1982b) and DiPietro (1983a, 1983b) described similar rocks in the Underhill Formation. They noted the lithic similarities between this unit and the lower Pinnacle Formation to the west and interpreted the Underhill Formation as an eastern facies equivalent of the Pinnacle Formation.

Quartz-laminated schist (CZuql): This lithology is a light-grey to green-grey, rusty weathering, medium-grained garnetiferous biotite-plagioclase-quartz-chlorite-muscovite schist containing minor amounts of magnetite. Biotite is common locally. Quartz laminations 3 to 4 mm thick are diagnostic of this unit and are separated by mica-rich layers. Tauvers (1982a, 1982b) suggested that these laminations are, in part, of primary sedimentary origin. At location H20 (Pl. 2), these laminations are parallel to the dominant schistosity and contain a strong lineation which suggests that the quartz laminations originated exclusively as a tectonic fabric.

Although the quartz-laminated schist has a distinct fabric, the unit has lithic similarities to both the Hoosac and Underhill Formations. It is possible that the quartz-laminated schist represents an intermediate facies between the Hoosac metawacke (CZhg) and the Underhill schists (CZu). The contacts between the quartz-laminated schist and surrounding units are not exposed, but are interpreted as faults because Hoosac metawackes to the west display a much weaker deformation fabric than the laminated schist.

The quartz-laminated schist described by Tauvers (1982a, 1982b) and DiPietro (1983a, 1983b) comprises the Jerusalem slice north of the Lincoln massif. In that area, the quartz-laminated schist is bounded to the west and east by the Jerusalem and Underhill thrusts zones, respectively, where mylonitic rocks are reported. DiPietro (1983a, 1983b) also described isolated localities of the quartz-laminated schist as outliers of the Jerusalem slice. In the South Lincoln area, the quartz-laminated schist is interpreted as a fault sliver that represents the southern extension of the Jerusalem and Underhill thrust zones.

The contact between the Hoosac and Underhill Formations is poorly exposed. At locations Q22, M22, and X21 (Pl. 2), the contact is defined by the west-to-east transition from a sequence of dominantly quartzo-feldspathic metawackes to muscovite-rich schists which occurs over a distance of several hundred meters. It should be understood that the contact between the Hoosac and Underhill Formations is highly obscured by Paleozoic folds and shear zones.

Stanley and Ratcliffe (1985) interpret this boundary as a major fault zone. The present study does not support or refute this interpretation. A gradational contact is shown on the map to indicate the relatively gradual change in lithologies between what is called the Hoosac and Underhill Formations in the South Lincoln area. Resolution of the nature of this awaits further work to the north and south of the study area.

STRUCTURAL GEOLOGY

Introduction

The structural history of the Lincoln massif is complex. Grenvillian folds and foliation fabrics are poorly preserved within the massif. Cover rocks contain two Paleozoic fold generations and three generations of Paleozoic foliation. The Lincoln and Ripton anticlines are large-scale Paleozoic folds cored by the basement rocks. In the western part of the Lincoln massif (WLM), Paleozoic folds deform the map-scale layering and Paleozoic faults are absent. In contrast, several major Paleozoic thrust zones are recognized within and along the eastern margin of the eastern part of the Lincoln massif (ELM). These include the Cobb Hill, South Lincoln, and Underhill thrust zones. The Cobb Hill thrust zone displays a complex variety of rock fabrics related to progressive deformation of the granitic gneiss of the ELM during the Taconian orogeny. The South Lincoln and Underhill thrusts represent regionally continuous fault zones in the Eastern Sequence.

Western Part of the Lincoln Massif (WLM)

Grenvillian Structures

Grenville age folds have not been documented within the WLM. A Grenvillian fabric (Sg) is, however, commonly observed throughout much of the Mt. Holly Complex. The Grenvillian schistosity is defined by compositional layering within the quartzo-feldspathic (Ymhg) and biotite (Ymhbg) gneisses. Grenvillian-age compositional layers within the augen gneiss (Ymhag) are defined by the segregation of coarse microcline augen up to 1 cm in diameter within a finer grained quartz-feldspar-mica matrix. This layering is typically deformed into broad open folds by Paleozoic (F2) folds. The complex fold pattern observed within the WLM (Pl. 1) is likely due to an interference of Grenvillian and Paleozoic folding, although the dominant folding recognized is Paleozoic.

Paleozoic Structures

<u>S1</u>: The oldest Paleozoic schistosity (S1) is well-developed within the Pinnacle metawackes. The S1 schistosity is defined by 0.5 to 1 cm thick quartz-feldspar/mica laminations subparallel to bedding (S0) in the metawackes along the western limb of the Ripton anticline (Prahl, 1985). It is uncertain whether these laminations represent a primary compositional layering or a tectonic fabric in the rock. Micas are strongly aligned parallel to these laminations and are interpreted to be the product of a tectonic overprint.

F2/S2: The dominant deformation fabric observed within, and to the west of, the WLM is represented throughout the western part of the study area by gently plunging, open folds (F2) and an axial surface foliation (S2). F2 folds are commonly observed within the Pinnacle Formation. They deform the S0/S1 fabric into broad open folds plunging from 5 to 30 degrees north. The quartzite (Ymhgq) and microcline augen gneiss (Ymhga) units within the WLM outline these folds (Pl. 1) which plunge 28 degrees to the north (Prahl, 1985).

The dominant schistosity throughout the western part of the study area (S2) is parallel to the axial surfaces of F2 folds and is defined by the preferred orientation of micas in all units. This schistosity is well-developed in the cover rocks and commonly observed, although weakly developed, within the basement gneisses. The orientation of the S2 schistosity in the cover rocks west of the WLM strikes N11W and dips to the east at 40 degrees (Prahl, 1985).

F3/S3: Although crenulations that deform the dominant schistosity within the Pinnacle Formation are rare, their presence does indicate that post-F2/S2 deformation has affected the western part of the Lincoln massif in a minor way.

Ripton Anticline

The Ripton anticline is a broad, open, slightly overturned, north-plunging fold interpreted to have developed during D2 deformation with the dominant regional schistosity (S2) parallel to its axial surface. The eastern limb of the structure is covered but is interpreted to have a moderate dip compared to the western limb. Bedding within the Cheshire Formation along the western limb of the Ripton anticline is slightly overturned to vertical, dipping east 85 to 90 degrees (Prahl, 1985). Analysis of minor F2 folds in the core and along the western limb of the anticline suggest a moderate northward plunge (<30 degrees) for this structure.

Shear Zones

A strongly foliated, mica-rich zone is recognized by Prahl (1985) as a Paleozoic shear zone of minor displacement within the western part of the WLM. This steeply dipping, north-south trending zone represents the only known Paleozoic fault zone development within the WLM.

Eastern Part of the Lincoln Massif (ELM)

Grenvillian Structures

Grenvillian folds are rarely observed, although evidence for large-scale Grenvillian folds is represented by the large mass of interbedded blue quartzite (Ymhq) and tourmaline-chloritoid schist (Ymht) in the northwestern part of the ELM (Pl. 2, Loc. L12-M12, L13-M13). These broad folds are Grenvillian in age because they do not involve the basal unconformity or the overlying Pinnacle metawacke. Stereonet analysis of bedding and associated

mullion lineations indicates that the fold axis plunges 52 degrees to the southeast (S65E; DelloRusso, 1986a).

A Grenvillian schistosity (Sg) is common in outcrops displaying only a weak Paleozoic fabric and is defined by flattened and rodded quartz grains whereas the Paleozoic fabric is defined by the preferred orientation of micas. The Grenvillian fabric is randomly oriented throughout the ELM. Stereonet analysis indicates that the average schistosity is oriented approximately N22E 58E. This orientation is similar to that of the dominant (S2) Paleozoic schistosity in the gneiss which suggests that many of the Grenvillian structures may have been locally reoriented parallel to the Paleozoic schistosity during deformation. As a result, Grenvillian structures can be misinterpreted as Paleozoic in age.

Paleozoic Structures

In the Paleozoic rocks surrounding the ELM, a complex history of fold and foliation development is preserved. Only one generation of folds and two foliations are observed in the Western Sequence. In contrast, two fold generations and three foliations are observed within the Eastern Sequence.

Map-scale Paleozoic folds are not present within the basement rocks of the ELM, although a consistently-oriented foliation (S2) locally overprints the Grenvillian foliation (Sg) and correlates well in orientation with the dominant Paleozoic schistosity in the cover rocks. It appears that Paleozoic strain in the basement rocks has been accommodated largely by numerous narrow zones of ductile shear (Pl. 1) rather than by extensive folding.

<u>S1:</u> The S1 schistosity is essentially parallel to bedding in conglomerates and metawackes of the Pinnacle Formation. It is similar to the early schistosity recognized by Tauvers (1982a, 1982b) and DiPietro (1983a, 1983b) and could be related to bed-parallel shear developed during the early stages of shortening in the region. Folds of this generation are not recognized in the field area.

Due to the pervasive development of the dominant schistosity (S2), the S1 schistosity has been transposed along the eastern margin of the massif. A relict foliation is present, however, in microlithons defined by the dominant schistosity. This S1 foliation is interpreted to be equivalent to the bed-parallel schistosity observed in the Pinnacle Formation to the west, although this correlation is tenuous due to the lack of structural continuity across the massif.

<u>F2/S2:</u> The first generation of folds (F2) is defined by a folded, bed-parallel schistosity (S1) and/or compositional layering in the Pinnacle metawackes west of the ELM. The F2 folds have an associated axial planar schistosity (S2) that has an average orientation, west of the ELM, of N08E 71E. The folds are commonly tight to isoclinal and some display a chevron-like style. The limbs of these folds are typically rotated into alignment with the dominant schistosity such that the older S1 compositional layering is indistinguishable from the dominant schistosity (S2). The dominant schistosity (S2) east of the Mt. Holly Complex is isoclinally folded and is

oriented N06W 76E. F2 folds are rarely observed in the Eastern Sequence and are represented by folded quartz pods.

F3/S3: The second generation of folds is defined by tight to isoclinal minor folds, crenulations, and outcrop-scale open folds that deform the dominant schistosity (S2) east of the basement complex. These folds are not observed within, or along, the western margin of the ELM. Map-scale F3 folds are represented in the South Lincoln area (Pl. 2, Loc. C18-D18) where they result in a complex intercalation of the schists and metawacke. Detailed work by Strehle (1985; Strehle and Stanley, 1986) has also shown that crenulations of the same generation deform the mylonitic fabric along the South Lincoln thrust zone. The average orientation of axial surfaces to F3 folds is N18W 36W. Fold axes trend N25W and plunge 18 degrees and crenulation lineations trend N22W and plunge 07 degrees. The westerly dip of the F3 axial surfaces may reflect rotation on post-F3 thrust faults or younger folds. Alternatively, the folds may have developed in that orientation, reflecting the steep nature of the east limb of the Lincoln anticline.

The S3 schistosity is not well-developed at the outcrop, although micas oriented parallel to the axial plane of crenulations are observed in thin section. The S3 surface is defined by discontinuous shear bands associated with these crenulations.

Lincoln Anticline

The Lincoln anticline is one of the few map-scale folds in the region. It is described by Tauvers (1982a, 1982b) as a tight, overturned structure at the northern end of the ELM. To the south, in the area of this report, the bedding on the western limb is nearly vertical to overturned while the eastern limb dips eastward, subparallel to the dominant schistosity (S2). The anticlinal structure is more asymmetrical and upright compared to the north where it is strongly overturned (Pl. 3, cross-section B-B'). This structure is interpreted to have developed synchronously with the development of the dominant (S2) schistosity (Tauvers, 1982a, 1982b; DiPietro, 1983a, 1983b; DelloRusso, 1986a).

Parasitic folds on the western limb of the Lincoln anticline are associated with the development of a high angle shear zone, along the basement-cover boundary (WMZ; Pl. 3, cross-section B-B'). Correlative folds have been mapped by Tauvers (1982a, 1982b) to the north and also by Prahl (1985) along the western limb of the Ripton anticline. These F2 folds are interpreted to have developed synchronously with the Ripton and Lincoln anticlines.

To the west, the syncline which complements the Lincoln anticline is tight to isoclinal with a minor anticline in the hinge area (Pl. 3, cross-section D-D'). The hinge of this structure is well-constrained by bedrock exposure and is defined by the trace of the basal unconformity. This syncline is interpreted to plunge at a relatively shallow angle to the north (<35 degrees), although fold axis data are rare. It is continuous with the south-plunging Sugar Hill syncline south of the village of Ripton (Osberg, 1952; Doll and others, 1961).

Paleozoic Fault and Shear Zones

Western Marginal High Angle Shear Zone (WMZ)

A major shear zone is exposed along the western margin of the ELM, displaying fabrics that indicate east-side-down relative displacement (Pl. 2). This zone is bounded by the Mt. Holly gneiss but incorporates slices of the basal Pinnacle conglomerate. The rock in the zone is a strongly-sheared quartz-sericite schist that displays a complex microscopic foliation fabric with up to three generations of foliation. This rock type may have been produced as a result of shearing and retrogression of the Mt. Holly gneiss (Ymhg), the Pinnacle conglomerate (CZpbc), or both. The dominant schistosity in this zone is oriented N10E and is vertical. Several exposures along the WMZ exhibit well- developed S-C fabrics that clearly indicate an east-side-down displacement across the zone (Pl. 2, Locs. T10, E12-F12, E13-F13; DelloRusso, 1986a).

At location R10 (Pl. 2) along the western marginal shear zone, a large exposure of fragmented gneiss (fault breccia) occurs on the eastern side of a ridge of massive quartz-feldspar gneiss (Ymhg). Large, angular to subrounded clasts of weakly foliated gneiss are surrounded by a strongly recrystallized, but weakly foliated, matrix of fine-grained quartz, feldspar, and sericite. A similar fabric is also observed in an isolated exposure at location R13 (Pl. 2; DelloRusso, 1986a, p. 132). This rock type probably evolved in the same manner as the fragmented gneiss exposures in the Cobb Hill thrust zone (Pl. 4).

The angular shape of many of the gneiss clasts suggests that brittle deformation occurred in the zone. The weak foliation in the rock and the abundance of coarse magnetite porphyroblasts in the matrix indicates that extensive metamorphism post-dated the fragmentation of the gneiss. In thin section, the clast/matrix contacts are obscured by mica growth across them. The contacts are easily recognized, however, because the matrix is finer grained and mica-rich. An older foliation, which is absent in the matrix, is present in the clasts and is interpreted as the Grenvillian schistosity (Sg). This foliation is oriented differently from clast to clast, indicating that it predates the fragmentation of the rock. These relations suggest that the fault zone rock evolved during Paleozoic orogenesis.

The fabric observed at location R10 (Pl. 2) indicates that brittle deformation preceded the later ductile deformation. These observations are similar to those in the Cobb Hill thrust zone to the east. These relations suggest that the physical conditions of deformation correspond to the brittle-ductile transition for granitic rocks. Furthermore, the similarity in rock fabric between the two zones suggests that they underwent deformation during the same orogenic event (Taconian orogeny).

At location F12-F13 (Pl. 2) a zone of intensely sheared Pinnacle metawacke is exposed on the north bank of a stream and is adjacent to the road. This shear zone is interpreted as the northern extension of the WMZ. The zone is at least 15 m wide, although its margins are not exposed. Upstream, to the west, of this zone the rock displays a very planar foliation. In contrast,

the shear zone displays a strong S-C fabric which indicates east-side-down relative displacement across the zone. Outcrop of the Mt. Holly gneiss is observed approximately 50 m east of this locality indicating that the zone is adjacent to the basement-cover contact.

Some exposures along the WMZ preserve classic examples of the basal unconformity. The well-preserved character of the basement-cover contact along the western margin of this shear zone (Pl. 2, Loc. S10, T10), as well as in slivers along the zone, provide good stratigraphic top indicators that define a minor anticline along the western limb of the Lincoln anticline. The shear zone evolved from a parasitic fold on the western upright limb of the Lincoln anticline. The WMZ is thought to have formed as a moderately west-dipping fault due to large-scale flexural slip across the adjacent basement-cover boundary in response to the development of the map-scale folds across the Lincoln massif (Lincoln-Ripton anticlines and intervening syncline). Subsequent rotation of the zone occurred with the steepening of the western limb of the Lincoln anticline. The mechanism of faulting associated with this style of flexural folding has been described by several workers (Ramsey, 1962; Tauvers, 1982a; Brace, 1953). High angle reverse faults observed in the zone indicate that the WMZ is indeed related to the Paleozoic compression.

The mineralogy and brittle-ductile nature of fault-zone fabrics observed along the WMZ suggests that deformation occurred at greenschist facies conditions. This shear zone is continuous along strike for approximately 5.3 km which indicates that it is a major fault. The amount of displacement across the zone is unknown, but is thought to be on the order of several hundred meters.

The eastward-directed WMZ may represent a mechanism for creating a zone of weakness in the upright or overturned limb of a large scale fold, which might later evolve into a major, westward-directed thrust zone. Evidence for such a mechanism is easily obscured by subsequent thrust faults. DelloRusso (1986a; 1986b) suggested that the development of the Hinesburg thrust north of the study area may be related to a zone similar to the WMZ.

Rare, minor, westward-directed thrust faults transect this zone (Pl. 2, Loc. R10). Such zones, however, serve to document that the WMZ developed prior to westward-verging thrust zones along the western margin of the ELM.

Minor Thrust Faults

Two minor thrust faults are present within the western cover sequence at locations X09 and X10 (Pl. 2; Pl. 3, cross-section C-C') along the north bank of Alder Brook. At Location X10 (Pl. 2), a fault is oriented N12E 60E and displays strong down-dip lineations on a 2.5 cm thick quartz vein along the fault surface.

At location X09 (Pl. 2), distinct thrust fabrics are well-preserved along a discrete fault surface in the Pinnacle metawacke (CZpm). Here a well-developed fault zone cleavage in the upper plate has been rotated into the fault as a result of east-over-west displacement. A second fault surface is

located approximately 65 cm above the last fault. A thick quartz-pyrite vein displaying well-developed slickenlines plunging 58 degrees to the east occurs along the fault. This surface is folded and deformed by the fault zone cleavage associated with the younger, lower fault. These relations suggest that, at least on a small scale, thrust zones are progressively younger from east to west in the area.

These minor thrust faults truncate and therefore postdate the dominant S2 schistosity and related F2 folds. Cross-section C-C' (Pl. 3) shows these faults crosscutting the WMZ because the faults deform the S2 schistosity which developed synchronously with the WMZ. The displacement across these fault must have been less than 50 m because the rocks of the upper plate do not carry the older basement gneiss westward to these localities.

Cobb Hill Thrust Zone

Paleozoic shortening of rocks within the Mt. Holly Complex appears to have been accommodated largely by the development of numerous ductile shear zones (Pl. 1). Nearly all of these zones display east-over-west shear fabrics and are consistent with the thrust interpretations of faults in the eastern cover sequence. The homogeneous texture of the gneiss appears to play a large role in the recognition of Paleozoic folds in the ELM. All of the faults and shear zones, except minor brittle normal faults previously described, display some degree of ductile deformation.

The Cobb Hill thrust zone, a major zone of ductile deformation, is well-exposed along the western slope of Cobb Hill and the eastern flank of the Alder Brook Valley. Much of the entire width of the zone is exposed along the steep western slope of Cobb Hill over a distance of approximately 400 m. The margins of the zone are unexposed and poorly defined. The zone extends from Alder Brook upslope to the east to an elevation of 1800 ft above sea level where a break in the slope separates more massive gneiss to the east from the mylonitic rocks of the thrust zone. South of Cobb Hill, three, small, closely-spaced but critically instructive exposures define the southern extension of the zone (Pl. 2, Locs. V13, W13). The northern extension of the Cobb Hill thrust zone is poorly exposed and is represented by several shear zones mapped along the New Haven River (Pl. 2, Locs. B16, B17, C17).

The Cobb Hill thrust zone extends the length of the study area and may correlate with the regional thrust zones proposed in the Green Mountain massif south of the study area (Karabinos and Thompson, 1984). The extent of this zone north of the study area is presently unknown.

The relatively uniform composition of the gneiss throughout the Cobb Hill thrust zone provides an excellent field laboratory for the analysis of the fault zone evolution. The rock fabric is locally very complex and a variety of rock fabrics have evolved along this zone. S-C mylonites are abundant and consistently indicate east-over-west displacement. Cross-cutting fabrics are rarely observed but, in a few localities, these relationships clarify the complex evolution of the thrust zone. Fault-zone rocks include: phyllonite, (proto-, ortho-, and ultra-) mylonite (as defined by Wise and others, 1984), folded mylonite, fragmented mylonite, mafic slivers, and quartz-sericite

veins. The locations of exposures exhibiting these fabrics is given on Plate 4 (see also DelloRusso, 1986a).

Four distinct phases of fabric development are observed in the Cobb Hill thrust zone: 1) gneiss fragmentation and anastomosing mylonitic schistosity, 2) pervasive mylonitic schistosity, 3) folded mylonitic schistosity, and 4) shear bands and fragmented mylonite (DelloRusso and Stanley, 1986). Thick zones of sericite phyllonite and milky vein quartz are also associated with these deformation fabrics. The four phases of fabric development and their characteristic fault-zone rocks are documented in Plate 4. A summary of the evolution of the fault zone, based on observations from three closely-spaced outcrops (Pl. 2, Locs. V13, W13), is also shown on Plate 4. A summary of important characteristics used to interpret the evolution of the fault-zone rocks in the Cobb Hill thrust zone is presented below.

<u>Phase 1A - Fragmented gneiss:</u> At several exposures, angular to subangular clasts of weakly-deformed coarse-grained gneiss, ranging from 1 to 25 cm in diameter, are present in a fine-grained quartz-feldspar-sericite matrix (Pl. 2, Locs. V13, Q15). The matrix is weakly to moderately foliated and wraps around the gneiss clasts. In thin section, the foliation in the strongly recrystallized matrix locally overprints the clasts and clearly transects the matrix-clast contacts. Some clasts, however, are not deformed by the younger foliation. The composition of the clasts is identical to that of undeformed gneiss. As a result of the sericitization of feldspar grains, the matrix is more sericite-rich than the gneiss clasts. The gneiss fragments are always supported by the matrix with elongate clasts aligned parallel to the foliation of the matrix. In thin sections where the matrix does not display a well-developed foliation, the clast-matrix contacts are gradational as a result of post-tectonic metamorphism. These "brecciated" zones are typically bounded by relatively massive granitic gneiss, although the actual contacts or transition zones are not exposed. The presence of gneiss clasts in this protomylonitic fabric suggests that fracture and cataclasis were the dominant deformation mechanisms during the early history of deformation in the fault zone.

<u>Phase 1B - Anastomosing mylonitic schistosity:</u> This fabric is best developed at location V13 (Pl. 2) where large, angular blocks of weakly deformed granitic gneiss are preserved within a mylonitic matrix. This fabric is analogous to the "brecciated" fabric previously described. Evidence for extensive retrogression of feldspar is not observed, however, in the matrix of the rock. Unlike the "brecciated" fabric (Phase 1A), the matrix surrounding these clasts is rich in fine-grained recrystallized quartz and feldspar. This fabric suggests that ductile deformation, rather than cataclasis, may have dominated the early-stage fragmentation of the gneiss.

<u>Phase 2 - Pervasive mylonitic schistosity:</u> This fabric is common throughout the Cobb Hill thrust zone and throughout some mylonitic zones to the west. Examples of this fabric are exposed at locations Q12, Q15, V13, S14 and W13 (Pl. 2) where quartz and feldspar are strongly recrystallized and form a very planar mylonitic fabric with well-developed quartz rods (ultramylonite of Wise and others, 1984). Isolated porphyroclasts display a strong and consistent asymmetry, indicating an east-over-west sense of shear along the fault zone. Where mica is abundant, a fine-scale S-C fabric is observed which

also indicates an east-over-west shear sense (Lister and Snoke, 1984; Simpson and Schmid, 1983). The pervasive mylonitic schistosity is uniformly finer grained, has rarer porphyroclasts, and represents a more advanced stage of the deformation in the shear zone than the Phase 1 fabrics.

Phase 3 - Folded mylonitic schistosity: Folded mylonitic schistosity is only observed at location V13 (Pl. 2) in close proximity to the fault fabrics of Phases 1 and 2. In this outcrop, the pervasive mylonitic schistosity is strongly deformed by 1 to 2 cm wavelength, asymmetrical folds with a consistent clockwise (looking east) asymmetry. The axes of these folds are subparallel to quartz rod and mineral lineations observed throughout the Cobb Hill thrust zone (Pl. 4, Fig. 16). This suggests that the folds have been rotated into near parallelism with the transport lineation in the zone. Folds displaying a sheath form, however, have not been observed. The reclined folds are believed to represent a transient phase between the earlier mylonitic fabric and the subsequent development of shear bands along the attenuated fold limbs.

Phase 4 - Shear bands and Fragmented mylonite: At location V13 (Pl. 2), shear bands truncate the attenuated limbs of the folded mylonitic schistosity (Phase 3 fabric). These shear bands develop as a result of continued deformation in the shear zone and represent a more stable fabric than the mylonitic schistosity (Sm foliation; White and others, 1980; Berthe and others, 1979). These bands fragment the folded mylonitic fabric and are defined by thin, very planar, sericite-rich zones. The bands dip to the east at a slightly shallower angle than the older mylonitic foliation (Pl. 4, Fig. 16).

As a result of fold and shear band development, some exposures contain fragments of mylonitic gneiss in a sparse matrix of sericite and fine-grained, recrystallized quartz (Pl. 2, Locs. V13, S14, Q15; see also Jegouzo, 1980). In hand specimen, the clast boundaries and the folded mylonitic fabric within them are obvious. In thin section, however, the clast-matrix contacts appear gradational because the recrystallized grains in the clasts and matrix are the same size. Due to the abundance of clasts in the rock, most exposures appear to have a relatively massive texture. A strong mineral lineation within the matrix foliation is inferred to be parallel to the transport direction in the thrust zone. Mylonitic clasts are stretched with their long axes oriented parallel to this lineation. Although this fabric has an apparent similarity to the fragmented gneiss of Phase 1, it differs in that the clasts contain a Paleozoic mylonitic foliation and not a Grenvillian fabric. This rock type is far more evolved than the fabric of Phase 1 and should not be confused with it.

Phyllonite: Two prominent phyllonite zones occur in the Cobb Hill thrust zone at locations S14 and V13 (Pl. 2). At location S14, continuous outcrop in the stream bed exposes a 5 m thick zone of sericite-carbonate phyllonite. The lower contact of the zone is sharp and the rock below this zone is a fine-grained massive granulite. A polished hand specimen of the rock, however, reveals that it is composed of fragments of mylonite. Although the upper contact is poorly exposed, the overlying gneiss displays a strong mylonitic fabric (Phase 2). S-C fabrics and asymmetric quartz-fiber growths on

magnetite porphyroblasts display a consistent asymmetry, indicating an east-over-west shear sense across the zone. Above this zone, the mylonitic gneiss displays a pervasive fabric with rare porphyroclasts. The S-C fabric of the mylonite and the asymmetry of porphyroclasts indicate east-over-west thrust displacement.

At location V13 (Pl. 2), Phase 1, 2, 3, and 4 rock fabrics are in close proximity to a 2 m thick zone of milky vein quartz and sericite phyllonite. The sericite phyllonite is clearly the product of gneiss alteration, because pods of gneiss in the quartz vein are surrounded by sericite. The gneiss is also strongly sericitized along its gradational contact with the phyllonite. Sericite pods are flattened and the gneiss fragments are stretched parallel to the mineral lineation in the Cobb Hill thrust zone.

The close association of these phyllonite exposures with fragmented mylonites and sericite-rich shear bands (Phase 4 fabrics) may indicate that the phyllonites also developed during the later stages of the fault zone evolution.

<u>Conditions of deformation:</u> The presence of an amphibolite sliver containing blue-green amphibole, epidote, and biotite (DelloRusso, 1986a; sample 280-2) and the presence of mylonites containing recrystallized feldspar within the Cobb Hill thrust zone indicates that the conditions of deformation reached minimum metamorphic conditions of the biotite zone/upper greenschist facies.

The coexistence of both fragmented gneiss (fault breccia) and mylonites suggests that deformation occurred in the brittle-ductile transition. The nature of the brittle-ductile transition in granitic rocks has been examined by numerous workers (Sibson, 1977; Tullis and Yund, 1977; Mitra, 1984; Jegouzo, 1980; Watts and Williams, 1979). Observations from field and experimental examples have defined the range of approximate physical conditions for the brittle-ductile transition in granitic rocks. Temperatures range from 250 to 400 degrees C and pressures range from 2.7 to 4.0 kbar. These values represent greenschist facies metamorphism. These pressures would correspond to depths between 9 and 12 km below the earth's surface.

At conditions of biotite grade metamorphism, the seemingly brittle fabric recorded in the protomylonites previously described (Phases 1 and 4) might not be expected to develop (Tullis and Yund, 1977; Sibson, 1977). The fragmented appearance of the rock, however, suggests that brittle mechanisms were dominant during an early stage of deformation. Although temperature is an important limiting variable that determines whether a rock deforms in a ductile or brittle manner, other such variables as strain rate, grain size, rock composition, and fluid interaction become very important when the temperature of deformation is within the brittle-ductile transition. Consequently, the development of fragmented gneiss (Phase 1) may have resulted from any one or a combination of the following:

1. The lack of a pervasive metamorphic fluid during initial deformation of the gneiss would preclude any reaction-enhanced softening (White and Knipe,

1976) or hydrolytic weakening (White and others, 1980). These processes would otherwise allow the rock to deform ductilely.

2. A high strain rate during deformation would allow the gneiss to deform by fracturing if the recovery rate of crystal-plastic mechanisms was exceeded (Mitra, 1980; Wise and others, 1984).

3. The coarse grain size of the gneiss would also allow brittle deformation to dominate over crystal-plastic mechanisms (Mitra, 1980, 1984a, 1984b).

4. Under conditions in the brittle-ductile transition, feldspar may deform in a brittle manner, whereas quartz would be dynamically recrystallized (Sibson, 1977). Because the granitic gneiss (Ymhg) is rich in feldspar, it would mainly deform by fracture and cataclasis.

Displacement: There is little data on which to base quantitative estimates of displacement across the Cobb Hill thrust zone. Considering the length and thickness of the zone, as well as the complex array of fault rocks within it, displacement on the order of several kilometers is not unreasonable.

Eastern Limb of the Lincoln Anticline

The South Lincoln and Underhill thrust zones are regionally continuous fault zones which may be typical of the eastern cover sequence. Several generations of thrusts have recently been documented east of the study area by Lapp (1986; Lapp and Stanley, 1986) and O'Loughlin (1986; O'Loughlin and Stanley, 1986). Their work illustrates an increase in the concentration of thrust zones eastward.

Extensive Paleozoic shortening has been accommodated by faults, shear zones, and folds in the schists and metawackes of the Hoosac and Underhill Formations. Many outcrops have at least some minor areas of intense schistosity that defines a shear zone. Although, few of these narrow zones of intense shear can be traced along strike, they likely form an anastomosing fabric of small shear zones which pervade the entire sequence. These zones, along with at least two zones of major displacement, are responsible for the lack of recognizable map-scale folds in these rocks. The contact relations between major lithic sequences are also strongly obscured by synmetamorphic faults. Lenses of schist in metawacke, or vice-versa, are common and may be the result of primary deposition, isoclinal folds, faults or any combination of these events. Due to the lack of continuous exposure in the Eastern Sequence, mapped fault zones are restricted to those zones that contain slivers of the granitic gneiss of the Mt. Holly Complex (Ymhg). Narrow zones displaying an intense shear fabric that can be traced out along strike are shown on the map (Pl. 1), because they represent zones of probable large displacement.

Underhill Thrust Zone

The map trace of the Underhill and Jerusalem thrusts of Tauvers (1982a, 1982b) has been reinterpreted by recent detailed mapping in the South Lincoln

area. The thrust exposed at the South Lincoln bridge, which was originally described by Tauvers (1982a, 1982b) as the Underhill thrust, has been renamed the South Lincoln thrust due to the presence of another major thrust zone east of the South Lincoln locality. This new zone is correlated with the Underhill thrust mapped by Tauvers (1982a, 1982b) and DiPietro (1983a, 1983b) north of the study area. The thrust at South Lincoln is now shown to merge with the Jerusalem thrust, which Tauvers (1982a, 1982b) and DiPietro (1983a, 1983b) mapped just west of the Underhill thrust zone (Pl. 1).

The Underhill thrust zone is defined at locations H20 and X20 (Pl. 2). At location H20 outcrops of the quartz-laminated schist (CZuql) are present. This lithology was described by Tauvers (1982a, 1982b) in the Jerusalem slice, north of the study area. The quartz-laminated schist is strongly sheared and typically displays numerous, discontinuous, narrow, strongly foliated zones with interlayered patches of silvery Underhill schist. Discrete faults or shear zones are impossible to trace, but the entire zone has a significant amount of distributed shear across it because of the abundance of asymmetrical structures and small-scale truncations. Directly to the west, at location H19 (Pl. 2), outcrops of Hoosac quartz-feldspar metawacke display tight folds of the S2 schistosity but show no evidence of intense shear. The quartz-laminated schist at this location is interpreted to be part of the Jerusalem slice which was displaced westward along with movement on the Underhill thrust.

Compelling evidence for the existance of the Underhill thrust zone is observed in the southeastern part of the study area at location X20 (Pl. 2). Several elongate exposures of quartz-feldspar gneiss (Ymhg) are bounded by nearby outcrops of biotite metawacke (CZhbg) and feldspathic conglomerate (CZhbc) of the Hoosac Formation (DelloRusso, 1986a). Two separate slivers of gneiss (Ymhg) are defined along this thrust zone. The easternmost sliver is well exposed in several outcrops over a distance of approximately 250 m along strike. The rock weathers light-grey to white in contrast to the darker grey and rusty appearance of the Hoosac metawacke. The gneiss displays a pervasive mylonitic fabric oriented N04W 74E with a strong downdip quartz lineation trending S58E and plunging 67 degrees. Comparison of modal analyses of these rocks with samples of the Mt. Holly gneiss to the west confirms their compositional similarity (DelloRusso, 1986a; Appendix B, note that the gneiss contains potassium feldspar, whereas the metawacke contains only plagioclase feldspar).

A small quartz-rich mylonitic gneiss outcrop occurs about 100 m southwest of the above mentioned outcrop of gneiss. The strongly foliated, mylonitic fabric in this granitic rock indicates that it is also a basement sliver. This gneiss sliver cannot be traced along strike because of a lack of outcrop. Fine-grained conglomerate is found approximately 30 m east of the gneiss sliver. The conglomerate is interpreted to be in depositional contact with the gneiss, although the actual contact is covered. Just to the northwest of this sliver (approx. 80 m N20W) is an outcrop of the Hoosac biotite metawacke (CZhbg). These slivers possibly represent topographic highs on the erosional surface of the basement which were sliced off during thrust faulting.

The presence of basement slivers in the Eastern Sequence indicates a significant amount of displacement across this fault zone, although the exact amount cannot be determined. Stanley and Ratcliffe (1985) suggest displacements on the order of 5 to 15 km for a similar zone in Massachusetts.

South Lincoln Thrust Zone

The South Lincoln thrust zone was first described by Tauvers (1982a, 1982b) and later analyzed by Strehle (1985; Strehle and Stanley, 1986) at the South Lincoln bridge exposure (Pl. 2, Loc. D19). At this locality, a sliver of mylonitic Mt. Holly gneiss (Ymhg) overlies strongly foliated chloritic schist containing carbonate pods (CZhms) and biotite-rich schistose metawacke (CZhbg) of the Hoosac Formation. The fault contact is exposed on the east bank of the river, directly below the bridge and extends southward across the river, where it is exposed beneath the water's surface. The mylonitic schistosity in the gneiss is oriented N14E 60E and displays a strong quartz elongation lineation trending S64E and plunging 52 degrees. Directly overlying the gneiss is coarse cobble conglomerate of the basal Hoosac Formation (CZhbc) which displays stretched cobbles up to 30 cm in length. The cobbles are composed largely of quartzite and rarely of gneiss. Detailed hand specimen and petrographic analyses from this exposure have clearly documented the east-over-west sense of displacement across the thrust zone (Strehle, 1985; Strehle and Stanley, 1986).

To the south at locations H20, I20, and J20 (Pl. 2) a narrow zone of highly sheared rusty schist is exposed for approximately 400 m. This zone is bounded by light-grey schistose metawacke. This shear zone is interpreted to represent the southern extension of the fault exposed at the South Lincoln bridge (Pl. 2, Loc. D19).

Due south along strike, at location L19 (Pl. 2), a narrow, rusty, schistose zone exhibiting an intense shear fabric is exposed along the top of a 50 m long outcrop of schistose metawacke. This zone exhibits an east-over-west S-C shear fabric and is correlated with the shear zone to the north. The location of this zone farther south is not clear due to the lack of bedrock exposure between this locality and the exposures of gneiss slivers to the south.

In the southeastern part of the map area, small exposures of Mt. Holly gneiss (Ymhg) occur at locations X18 and X19 (Pl. 2). These outcrops are strongly foliated and are bounded by nearby outcrops of biotite-rich metawacke of the Hoosac Formation. They have a similar composition and fabric to the gneiss slivers along the Underhill thrust zone directly to the east. These slivers of basement gneiss are interpreted to have been incorporated into the cover sequence along a major thrust zone located near the basement-cover boundary. The slivers are, however, not everywhere coincident with that contact.

Displacement across the South Lincoln thrust zone is estimated to be similar to or less than the displacement across the Underhill thrust zone.

Joints

The analysis of joints, recorded throughout the study area, indicates that that many joints in the Mt. Holly Complex (N11E 57E) follow the trend of the Paleozoic foliation (S2, N05E 59E; DelloRusso, 1986a). A second, poorly-developed set of fractures, is oriented at a high angle to the Paleozoic fabric. These joints are similar in orientation to the brittle normal faults. This similarity suggests a genetic relationship between them.

Brittle Normal Faults

Brittle normal faults, although not abundant, are locally observed throughout the study area. These faults truncate all other rock fabrics observed in the outcrop and are clearly younger than the Paleozoic fabric. The majority of faults trend north-south and dip steeply eastward suggesting that their orientation may be influenced by the Paleozoic fabric. The displacement across most of these faults is limited to 1 to 3 cm (Pl. 2, Locs. L19, N17, Q13, Q12). Some of these faults display slickenline grooves, whereas others appear as joints which may have experienced minor adjustment. A few faults display a strong shear fabric and may have dip-slip displacements (up to 1 m; Pl. 2, Loc. M13). The lack of fault breccia or gouge along these faults is consistent with minor displacements across them. Their relation to known Mesozoic faults in the region and the Mesozoic rifting of the Atlantic ocean is uncertain but their orientation and fabric is compatible with such an event (Stanley, 1980).

Comparison of the ELM and WLM

As a result of the work outlined in this report, it is evident that there are major stratigraphic and structural contrasts between the eastern and western parts of the Lincoln massif. The stratigraphy of the WLM is dominated by distinctly layered gneisses interpreted as a pre-Grenvillian metasedimentary sequence. In contrast, the dominant lithology of the ELM is a relatively massive granitic gneiss thought to be of igneous origin. The presence of pre-Grenvillian metasedimentary rocks within the ELM, however, suggests that the ELM and WLM are not entirely different terranes.

The most obvious contrast between the ELM and WLM is the relative extent and style of Paleozoic deformation observed within the basement rocks. The ELM is strongly deformed internally by extensive Paleozoic ductile fault zones, whereas the WLM displays a coherent sequence of map-scale open folds with little or no evidence for faults. There is no evidence for a major fault zone separating the ELM and WLM. The deformational fabrics observed within the massif become progressively more intense from west to east across the area. This is also consistent with the dominant Paleozoic metamorphic fabric in the area. Furthermore, the lack of Paleozoic faults in the WLM suggests that folding predates the development of thrust zones in the basement rocks.

METAMORPHISM

Introduction

The minimum metamorphic conditions of Grenvillian and Paleozoic metamorphic events in the study area have been determined through detailed petrographic analysis of metawackes, aluminous schists, and mafic schists. Analysis of the metamorphic history of these rocks is done in concert with the structural fabric, using correlations of the foliation throughout the area as a framework in which to compare changes in metamorphic grade from west to east.

Grenvillian Metamorphism

The compositional layering observed in the layered mafic schist (Ymhml) of the eastern part of the Lincoln massif (ELM) is interpreted as a Grenvillian fabric (DelloRusso, 1986a). In thin section, this layering commonly contains a relict assemblage of coarse-grained plagioclase (andesine), blue-green amphibole, and garnet. The garnet-bearing mafic schist is only observed at location L12 (Pl. 2). Relict Grenvillian garnet is also observed in weakly deformed granitic gneiss in the southwestern part of the ELM (Pl. 2, Loc. FF7). This mineralogy indicates a minimum metamorphic grade for Grenvillian metamorphism at the garnet zone, which is equivalent to the epidote-amphibolite facies assemblage in the mafic schist.

Although relict pyroxene is not observed in the mafic rocks of the eastern part of the Lincoln massif, pyroxene has been reported by Prahl (1985) from the western part of the Lincoln massif (WLM). This relationship suggests that Grenvillian metamorphic conditions reached a much higher grade than is presently evident in the ELM (minimum amphibolite facies). Furthermore, this indicates that Paleozoic retrogression of the Grenvillian assemblages was more extensive in the ELM than in the WLM.

Paleozoic Metamorphism

Pre-Peak-M1 Metamorphism

Along the western margin of the WLM there is limited data relating to metamorphism associated with the S1 fabric. Muscovite aligned parallel to the compositional layering in metawacke of the Pinnacle Formation suggests a low grade of metamorphism (lower greenschist facies) associated with that fabric.

There is little evidence for extensive mineral growth prior to the development of the dominant schistosity (S2) along the western margin of the ELM. In some outcrops, however, muscovite is parallel to the compositional layering (S1) in the Pinnacle metawacke (CZpm).

A pre-S2 (S1) fabric is not observed in metawackes of the Eastern Sequence and can only be defined in thin sections of the mica-rich schists. The S1 schistosity is well-preserved in the schist of the Underhill Formation where it is defined by the alignment of sericite and quartz in the microlithons

between the S2 schistosity. Although these minerals do indicate recrystallization at greenschist facies conditions during the formation of S1, they do not provide evidence for more specific temperatures and pressures.

In mafic schists (CZhms) of the Eastern Sequence, evidence for pre-S2 mineral growth is preserved as relic, opaque-riddled cores in blue-green hornblende. These may represent an earlier generation of amphibole growth which would suggest metamorphism at greenschist facies conditions. We believe that this recrystallization occurred during the early stages of M1 rather than during a separate metamorphic event.

M1 Metamorphism

The dominant metamorphic fabric of the region (M1), which is apparent in all of the rocks within the study area, is associated with the D2 deformational fabric. West of the WLM, biotite and chlorite are commonly observed within the dominant foliation (S2) in metawackes of the Pinnacle Formation. Biotite is rare within the Fairfield Pond Formation because of its bulk composition. Biotite is observed within the dominant Paleozoic schistosity (S2) in the basement gneisses of the WLM. Grenvillian hornblende in amphibolite (Ymha) of the WLM is altered to biotite and chlorite as a result of Paleozoic metamorphism (M1).

The rocks of the Pinnacle Formation along the western margin of the ELM consist largely of quartz, feldspar, and sericite and contain only minor chlorite and biotite. As a result, these rocks are mineralogically insensitive to small changes in metamorphic conditions. Biotite is very rare but, where observed, it is unaltered and weakly aligned parallel to the dominant schistosity (S2). Chlorite, which is abundant locally, displays a similar relationship. Amphibole is not observed in chlorite-rich samples. Garnet was not found in the Western Sequence. Since rocks of similar composition in the Eastern Sequence do contain garnet, the absence of garnet in the Western Sequence indicates that the temperature and pressure during metamorphism in the west was lower that the temperature and pressure in the east. The occurrence of biotite parallel to the S2 schistosity indicates that minimum conditions of the biotite zone/greenschist facies were reached during the D2 deformation in the western part of the study area.

In the layered mafic schist (Ymhml) of the ELM, the plagioclase/amphibole segregation that defines the Grenvillian layering is transected by a single, well-developed Paleozoic foliation consisting of biotite, chlorite, albite, quartz, epidote, sphene, and opaques (DelloRusso, 1986a). In some samples, actinolite coexists with chlorite and altered, relict Grenvillian hornblende. This assemblage clearly retrogrades the earlier Grenvillian assemblage and indicates minimum metamorphic conditions of upper greenschist facies/biotite zone for the M1 metamorphic event.

The dominant Paleozoic foliation observed in the Mt. Holly Complex is correlated with the dominant schistosity (S2) in the Pinnacle metawacke to the west. The metamorphic grade associated with S2 in the western part of the ELM is consistent with that observed in the western cover sequence and is significantly lower than that observed in the Eastern Sequence.

In the Eastern Sequence, biotite is abundant in most metawacke samples, with garnet only occurring locally. Both of these minerals grew syntectonically with the S2 schistosity indicating that minimum metamorphic condition of the garnet zone were reached during that event. This is consistent with observations in the mafic schists of the Eastern Sequence.

Garnet is also commonly observed in the more aluminous schists of the Eastern Sequence. The garnet is syn- to post-kinematic with the S2/F2 event (M1) in the assemblage garnet, albite, quartz, and muscovite. Chloritized garnet porphyroblasts are common locally. In thin section, S3 crenulation cleavage is deflected around the garnet, whereas the pervasive schistosity (S2) is included in the porphyroblast. This texture indicates that metamorphic conditions at the garnet grade persisted for some time after the development of the S2 schistosity east of the Lincoln massif. Evidence for this sequence is lacking in the mafic schists.

In mafic schist samples from the Hoosac Formation (CZhms) the common assemblage associated with the dominant schistosity (S2) is blue-green hornblende, albite, biotite, chlorite, epidote, quartz, and opaques. A second, colorless amphibole coexists with blue-green hornblende in some mafic schist samples (Tauvers, 1982a, 1982b; DelloRusso, 1986a). Some blue-green amphibole grains exhibit ($\bar{1}01$) exsolution lamellae (DelloRusso, 1986a; Lapp, 1986). At location X20 (Pl. 2), garnet is also part of this assemblage. This mineral assemblage indicates minimum metamorphic conditions of the medium-pressure sub-facies series of the epidote-amphibolite facies (Miyashiro, 1973) for the dominant Paleozoic metamorphic event (M1) along the eastern margin of the Lincoln massif. The occurrence of kyanite, which lies within the dominant schistosity in the Mt. Abraham Schist (Cady and others, 1962; Lapp, 1986) to the east of the study area, also defines the medium-pressure subfacies series metamorphism for the dominant regional event.

Post-S2/M2? Metamorphism

Minor alteration of S2 biotite to chlorite has been observed in the western part of the Lincoln massif (Prahl, 1985). Similar post-S2 affects have been documented by Tauvers (1982a, 1982b) in the Lincoln area. These observations suggest that a very weak, post-M1 metamorphism occurred to the north and west of the WLM.

Post-S2 metamorphism or an S3 fabric has not been recognized in the Western Sequence along the western margin of the ELM. In the Eastern Sequence, however, significant syn-S3 mineral growth occurs.

In mafic schist (CZhms) of the Eastern Sequence, evidence for post-S2 mineral growth is largely represented by chlorite that has grown from biotite, hornblende, and garnet. Chlorite also occurs as randomly-oriented, crosscutting laths. The lack of extensive retrogression of the M1 assemblage and absence of strong deformation of the S2 fabric indicates that post M1 metamorphism was minimal and that the growth of chlorite is likely a late-stage M1 retrogressive effect rather than a distinct structural and metamorphic event.

Alteration of M1/S2 biotite to chlorite and the development of new biotite grains are common in the Hoosac metawacke (CZhg). In thin section, the randomly oriented biotite grains include the S2 surface defined by quartz, albite, and muscovite and appear undeformed by crenulations which deform the S2 foliation. Chlorite is most abundant near discontinuous, late crenulation bands. From these observations, it is evident that some post M1/S2 metamorphism occurred during the S3/F3 deformation.

Based on the above mineral assemblage, the minimum metamorphic conditions associated with S3/F3 deformation are of the biotite zone/greenschist facies, which is distinctly lower than the S2/F2 (M1) assemblage.

Summary

Evidence for proposed high-grade (amphibolite to granulite facies) Grenvillian metamorphism in the Mt. Holly Complex is lacking. Relict mineral assemblages indicate minimum Grenvillian metamorphic conditions of the epidote-amphibolite facies which is far below that observed in the Adirondack and Berkshire Grenvillian terranes. The lack of high-grade assemblages in the Lincoln massif is likely due to extensive Paleozoic retrogression of the basement complex.

The muscovite growth associated with the S1 schistosity observed in the cover rocks represents Paleozoic pre-peak M1 metamorphism at a minimum of biotite grade subfacies to the west and garnet grade subfacies to the east. This metamorphic fabric probably represents an early stage of the M1 metamorphic event.

It is evident that there is a distinct decrease in the Paleozoic metamorphic grade (M1) associated with the dominant schistosity (S2) from east to west across the massif. Garnet-bearing assemblages of the epidote-amphibolite facies/garnet zone in the Eastern Sequence indicate higher grade metamorphic conditions than the biotite-bearing assemblages of the greenschist facies/biotite zone in the Western Sequence. The nature of the transition from the biotite to garnet zone, which occurs within the Mt. Holly Complex, is obscure due to the lack of continuous exposure and the abundance of syn- to post-S2 shear zones which strongly transpose preexisting rock fabrics. The transition in metamorphic grade observed from east to west across the Lincoln massif is believed to be a gradual transition rather than an abrupt, fault-related one. In the Eastern Sequence, however, where syn-S2/M1 to pre-S3/M2 fault displacement is thought to be large, the transitions between assemblages of distinctly different metamorphic grade may be abrupt (for example, the Jerusalem thrust zone of Tauvers, 1982a, 1982b).

Evidence for the existence of a second, distinct metamorphic event (M2; Tauvers, 1982a, 1982b) is represented by chlorite and sericite associated with a crenulation cleavage (S3) and by randomly-oriented biotite in the Eastern Sequence. The extent of this event appears to be limited. More extensive post-M1 metamorphism has been documented east of the study area (Lapp, 1986; Lapp and Stanley, 1986; O'Loughlin, 1986; O'Loughlin and Stanley, 1986). F3/S3 fabrics are not well documented west of the Eastern Sequence in the

study area. It is possible that the S3 mineral growth may be related to a late-stage M1 event.

Laird and others (1984) have interpreted some of the polymetamorphic evidence and isotopic-age data in the eastern cover sequence to be the result of both the Taconian and Acadian orogenies. This same evidence, however, has been interpreted by Sutter and others (1985) to result from Taconian metamorphism and subsequent cooling. Resolution of this problem awaits further study.

EVOLUTION OF THE LINCOLN MASSIF

The synthesis of the Paleozoic depositional and tectonic history of the Lincoln massif is based on the information presented in the previous sections. Deformation and metamorphism associated with the Taconian orogeny has overprinted most of the primary features in these rocks. Detailed analysis of the rock fabric reveals the sequence and boundary conditions that were important in their evolution.

The Proterozoic history of the basement rocks is complex and greatly obscured by Paleozoic orogenesis. It is evident that at least part of the Mt. Holly Complex is of sedimentary origin. Some of these metasedimentary rocks mantle an older granitic gneiss of probable igneous origin. These rocks suffered intense deformation and metamorphism during the Grenvillian orogeny. Only minimal evidence of this deformation remains in the study area. Petrographic analysis indicates minimum metamorphic conditions of epidote-amphibolite facies were achieved during this event. Metamorphism probably reached a considerably higher grade based on regional relationships within other Grenville basement terranes unaffected by Paleozoic metamorphism (Wiener and others, 1984).

Numerous mafic dikes intruded the massif in post-Grenvillian time and may be associated with the Late Proterozoic rifting of the basement. The Late Proterozoic to Early Cambrian rifting of the basement complex resulted in the deposition of the cover sequence that now mantles the Mt. Holly Complex. The coarse clastics of the Pinnacle Formation represent proximal deposits derived from the adjacent basement rocks. Although the Hoosac and Underhill Formations contain some coarse clastic materials, they are thought to represent a more distal, eastern facies of the Pinnacle Formation because they contain an abundance of mica schist and mafic schist. Rocks of the Fairfield Pond and Cheshire Formations document the rift-drift transition along the western margin of Iapetus during the Late Proterozoic to Early Cambrian.

The deformational and metamorphic history of the area associated with the Middle Ordovician Taconian orogeny is also complex. The relationships between the major fold structures and fault zones that developed during the evolution of the Lincoln massif are illustrated in Figure 2 (A to D). The evolution of the foliation and metamorphic fabric observed is not illustrated in these diagrams but will be synthesized below.

Stage A

The basement begins to deform into broad open folds with minor shear zones developing along the basement-cover contact. The compositional layering and bedding plane schistosity (S1) in the cover rocks is progressively deformed from tight, upright folds in the east to broad, open folds in the west (F2). A pervasive axial plane foliation (S2) in the east is represented by a weak crenulation cleavage to the west. Foliation develops in the basement rocks. The metamorphic grade approaches the garnet grade east of the ELM and the biotite grade to the west (peak-M1).

Stage B (Syn-Post-Peak-M1 Metamorphism)

As the Lincoln anticline begins to develop, thrust zones (South Lincoln and Underhill thrust zones) become progressively better developed along the eastern limb of the structure, paralleling the basement-cover contact. These faults cut F2 folds structures and the dominant schistosity (S2). Along the steeper, western limb of the Lincoln anticline, parasitic folds begin to shear-out and develop into an eastward-verging thrust zone (Western Marginal Zone). Folding remains dominant in the Ripton anticline to the west as evidenced by broad, map-scale folding of the basement-cover unconformity. Deformation in the basement is represented by the continued development of a weak foliation. Fault zones within the basement are absent at this stage.

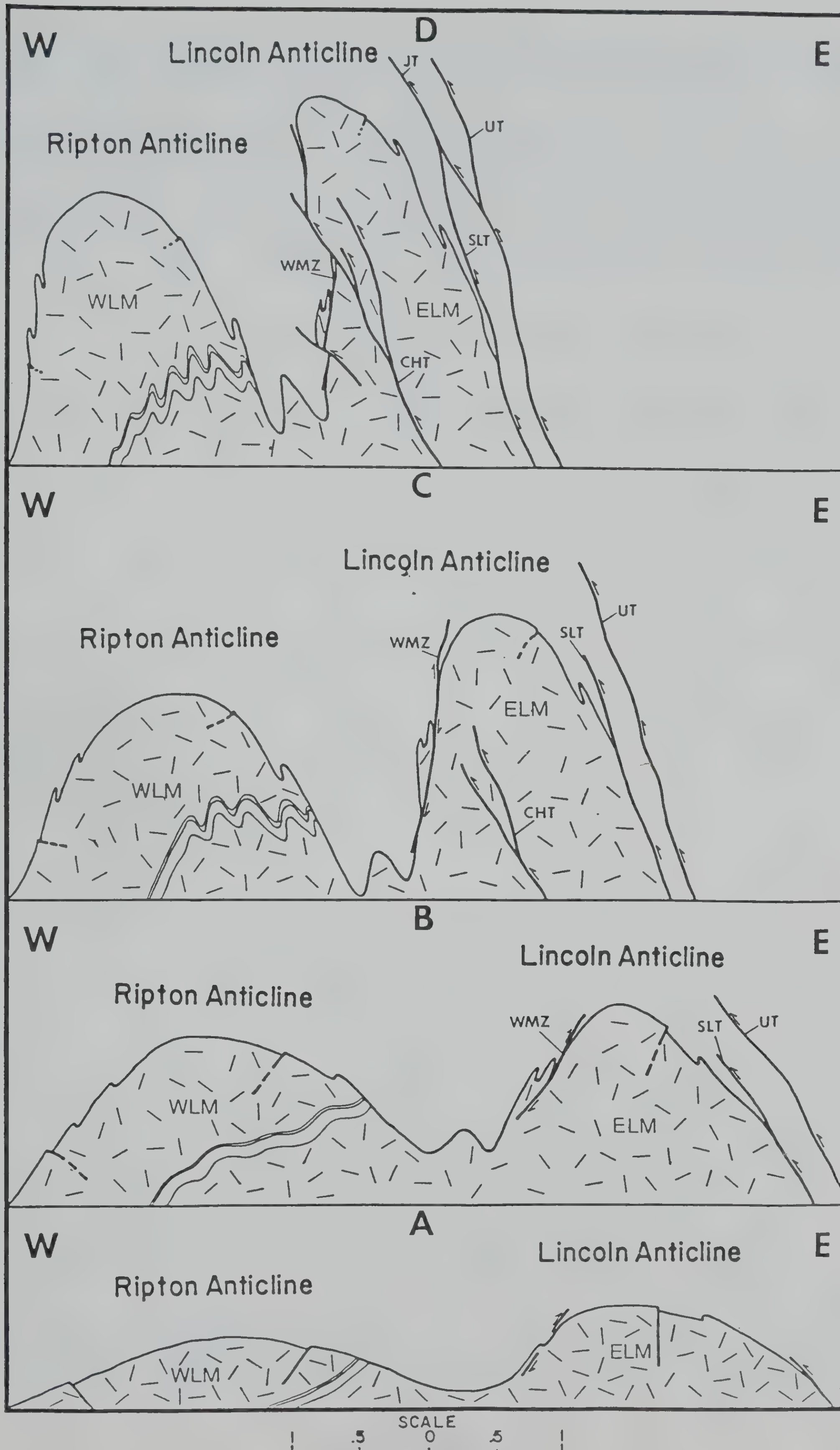
Stage C (Syn-Post-Peak-M1 Metamorphism)

The limbs of the Lincoln anticline continue to steepen in response to horizontal compression. Thrust zones east of the ELM steepen also, reflecting the orientation of the basement-cover contact. As these zones continue to rotate, with the limb of the anticline, they become sub-parallel to the dominant schistosity (S2). Ductile shear zones begin to develop from east to west within the ELM in response to the westward displacement of the Lincoln anticline (Cobb Hill thrust). In the western upright limb of the Lincoln anticline, displacement within the eastward-verging shear zone begins to die out, as the fold structure becomes overturned. To the west, the dominant schistosity (S2) becomes pervasive and is associated with chevron to isoclinal folds in the cover rocks and open folds within the basement of the WLM.

Stage D (Syn-Post-Peak-M1 Metamorphism)

As the Lincoln anticline becomes isoclinal and begins to overturn, shear zones in the ELM (Cobb Hill thrust) propagate westward and truncate the western limb of that structure. The eastern part of the ELM and its cover is displaced westward along the Cobb Hill thrust zone, with respect to the western part of the ELM and the WLM. To the east, steeply dipping thrust zones, which parallel the dominant schistosity, pervade the eastern cover sequence. Tight to isoclinal folds (F2) become prevalent to the west with well-developed folds in the basement rocks, while east-over-west ductile thrust zones pervade the ELM. Metamorphism reaches its peak at the garnet zone east of the ELM, decreasing in grade to the biotite zone to the west.

Figure 2. Schematic evolution of the Lincoln anticline (stages A to D). Stage A represents the early development of large scale basement folds, with minor shear zones developing on the limbs of the Lincoln anticline. Stage B illustrates the continued tightening of the Ripton and Lincoln anticlines, and the development of the western marginal shear zone (WMZ) as an eastward-verging thrust on the western limb of the Lincoln anticline. Westward-verging thrusts develop on the eastern limb of the Lincoln anticline represented by the South Lincoln (SLT), Underhill (UT) and Jerusalem thrusts (JT). During stage C, the western limb of the Lincoln anticline is steepened and, as a result, displacement along the WMZ dies out. Fault zones develop progressively westward across the ELM in association with the development of the Cobb Hill thrust zone (CHT). Shortening in the WLM is accommodated by folding. Stage D represents the present day cross-section. Note that a major zone of thrust faults separate the basement of the Lincoln massif from the eastern cover sequence. The CHT may project westward and transect the WMZ. Note the absence of fault zones in the WLM.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Subsequent shortening deforms the dominant schistosity in the eastern cover sequence into crenulations and open, east-facing folds (F3) which are consistent in orientation with the steep attitude of the basement-cover contact. Some syn- to post-D2 faults are folded, while others may have undergone renewed, although limited, movement. This deformation is weak to the west of the ELM and does not develop a recognizable fabric in the rocks. This event is associated with retrogression of the garnet grade assemblages east of the massif at minimum metamorphic conditions of the biotite zone (post peak-M1 or M2?) and may be related to the Acadian orogeny, although the precise age of this fabric is unknown.

REGIONAL CORRELATIONS AND IMPLICATIONS

Recent work based upon plate tectonic theory by numerous authors has resulted in a reinterpretation of many of the original ideas concerning the nature of the geologic history of Vermont. The interpretation that most of the rocks of eastern Vermont represent a complex sequence of lithotectonic assemblages rather than a simple east-dipping homoclinal sequence is based upon the discovery of numerous thrust zone within the serpentine belt (Stanley and Roy, 1982; Stanley and others, 1984) and along the eastern margin of the Berkshire and Green Mountain massifs (Stanley and Ratcliffe, 1983, 1985). Mapping within the Berkshire massif of western Massachusetts and the Green Mountain massif of Vermont has shown that the Grenvillian basement has been mobilized and broken up into numerous basement thrust slices (Ratcliffe, in Zen and others, 1983; Karabinos and Thompson, 1984; Ratcliffe and Burton pers. comm., 1986). Deep crustal seismic surveys also support these interpretations (Ando and others, 1983, 1984). Detailed mapping in this and adjacent areas provide a basis for testing the validity of the interpretations outlined above. The following discussion will focus on comparisons of our interpretations and conclusions with those of other workers in the region.

Stratigraphy

In the Mt. Holly Complex of the eastern part of the Lincoln massif (ELM), the granitic gneiss is considered to be an igneous intrusive. The gneisses in the western part of the Lincoln massif (WLM) are, however, thought to be metasedimentary rocks (Osberg, 1952; Prahl, 1985). The Grenvillian gneisses of the Green Mountain massif (GMM) of southern Vermont have also been interpreted as metasedimentary rocks because they contain marbles and quartzites. Granitic orthogneisses occur with paragneisses in the Adirondack massif of New York (Wiener and others, 1984). Grenvillian orthogneisses are also recognized within the Berkshire massif (Tyringham Gneiss of Ratcliffe and Zartman, 1976). Brace (1953) differentiated gneisses in the Rutland area of the Green Mountain massif (GMM) and suggested that some of these were of an igneous origin. Furthermore, some of these orthogneisses are interpreted as concordant sills which intruded a paragneiss sequence (Ratcliffe and Zartman, 1976). The existence, therefore, of both orthogneiss and paragneiss in the Mt. Holly Complex should be expected.

The metaquartzite (Ymhq) and tourmaline-chloritoid schist (Ymht) sequence of the ELM is correlative with similar lithologies described by Brace (1953) in the Rutland area as well as with quartzites described by Prahl (1985) in

the WLM. This correlation suggests that these massifs originated from the same basement terrane, although gneisses associated with these metasedimentary rocks may be genetically different.

In the Western Sequence, the rocks of the Pinnacle, Fairfield Pond, and Cheshire Formations of the western part of the study area correlate well with lithologies described by Tauvers (1982a, 1982b) and DiPietro (1983a, 1983b) to the north and by Osberg (1952) to the south. The basement-cover contact throughout this region is largely unfaulted. Recent mapping by Karabinos and Thompson (1984), however, has revealed that this boundary, along the northern part of the GMM, may involve Taconian thrust faults.

Regional stratigraphic correlations in the Eastern Sequence are hampered by limited exposure and the lack of detailed mapping in adjacent areas. Cady and others (1962) mapped the "Hoosac Formation" of Doll and others (1961) as the Pinnacle Formation because both units contain abundant metawacke. It is evident that the initial interpretation was that the Pinnacle Formation extended eastward across the ELM undisturbed by major faulting. Indeed there are similarities between the two sequences and they are interpreted to be equivalent lateral facies. The "Hoosac Formation", however, is compositionally different from the Pinnacle Formation because it contains biotite schist and mafic schists which are not present in the Western Sequence. The discovery of major thrust zones within and east of the ELM indicates that the Eastern Sequence originated from a more easterly position with respect to the Western Sequence.

Correlations along strike are tenuous in the Eastern Sequence. Conglomerates exposed on Mt. Holly gneiss slivers, however, represent the basal conglomerate overlying the Mt. Holly Complex elsewhere and are similar in morphology to the conglomerate of the Western Sequence. Conglomerate in the Hoosac Formation correlates with conglomerates in the Tyson Formation of the Plymouth area, Vermont (Chang and others, 1965). Osberg (1952) also describes similar conglomerates, along the eastern margin of the ELM, south of this study.

Correlation between the Hoosac and Underhill Formations, based on the distribution of lithologies, suggests that the Hoosac Formation grades into the Underhill Formation of Tauvers (1982a, 1982b) to the north as well as to the east of this study area. Previously, the Hoosac Formation was interpreted to terminate just north of South Lincoln, Vermont (Tauvers, 1982a, 1982b; Doll and others, 1961; Cady and others, 1962). Inasmuch as there is little or no bedrock exposure in this area, the nature of the transition between the Hoosac Formation and the Underhill Formation is not well known. In this study, the transition represents a change in lithofacies that does not necessitate a truncation of the basal stratigraphy in the South Lincoln area. Evidence from this study indicates that north-south stratigraphic continuity does exist in this region between the Hoosac and Underhill Formations. In light of this, a re-examination of the stratigraphic nomenclature may be in order. A much more detailed and consistent redefinition of these formations is necessary throughout central Vermont.

Structural Geology

Comparisons of the Paleozoic structural and metamorphic fabric from the study area, with those of surrounding areas, show a great deal of consistency. Detailed work by Tauvers (1982a, 1982b) and DiPietro (1983a, 1983b) to the north revealed two major fold generations (F1/S1 and F2/S2) that are similar to the generations described in this study. F2/S2 fold generations of Tauvers (1982a, 1982b) and DiPietro (1983a, 1983b) correlate well with the F2/S2 and F3/S3 fold generations of this study. DiPietro (1983a, 1983b) also recognized a bedding-plane schistosity, that is analogous to the S1 schistosity of this study. Tauvers (1982a, 1982b) described minor folding of the basement-cover unconformity along the western limb of the Lincoln anticline and related it to his second generation of folding, post-dating development of the Lincoln anticline. We disagree with this interpretation, because structural relations in this study suggest that parasitic folds on the western limb of the Lincoln anticline developed synchronously with the anticline. The lack of detailed systematic analysis of structural data directly south of the study area precludes a comparison of the structural history there.

Major fault zones recognized by Tauvers (1982a, 1982b) in the Lincoln area correlate to thrust zones documented in this study along the eastern limb of the Lincoln anticline. Tauvers (1982a, 1982b) proposed that a single thrust zone extended south of South Lincoln along the basement-cover contact. The recognition of numerous Mt.Holly gneiss slivers in the Eastern Sequence in this study indicates that a major thrust zone does exist there (South Lincoln thrust zone). Furthermore, it is apparent that thrust zones are not confined to the basement-cover boundary in this study area and that several zones may exist further east (Underhill thrust zone). Detailed mapping by Lapp (1986; Lapp and Stanley, 1986) and O'Loughlin (1986; O'Loughlin and Stanley, 1986) to the east has revealed that thrust zones are abundant there. Recent work by Karabinos (1984) has revealed major thrust zones along the eastern margin of the Green Mountain massif (GMM) in southern Vermont. These zones correlate with the Middlefield thrust zone in Massachusetts which separates the allochthonous Eastern Sequence from the basement gneisses of the Berkshire massif (Norton, 1971; Ratcliffe and Hatch, 1979). As a result, a regionally continuous zone of thrust faults is mapped along the eastern margin of the Berkshire, Green Mountain and Lincoln massifs (Stanley and Ratcliffe, 1985).

A minor shear zone described by Tauvers (1982a, 1982b) in the Mt. Holly Complex, west of South Lincoln, correlates with the northern extension of the Cobb Hill thrust zone. This relationship suggests that this zone may extend northward into the cover sequence. This zone is not yet recognized within the southern part of the Lincoln massif but may correlate to major thrust zones proposed by Karabinos and Thompson (1984) within the northern part of the GMM.

The western marginal shear zone (WMZ) is not recognized beyond this study area. This zone may extend northward along the western limb of the Lincoln anticline. Its relationship to the Hinesburg thrust is highly interpretive but correlates well with the model proposed by Tauvers (1982a, 1982b).

Metamorphism

A comparison of the metamorphic history of the study area with that of Tauvers (1982a, 1982b) also illustrates consistency between these two areas. Petrographic analyses from both studies indicate minimum metamorphic conditions of epidote-amphibolite facies in the east and biotite zone/ greenschist facies to the west associated with the first generation fold structures (M1). Existing radiometric dates indicate that this event is related to the Taconian orogeny (Cady, 1969; Laird and Albee, 1981; Laird and others, 1984; Sutter and others, 1985). The M1 metamorphic gradient increases to the east where a kyanite-chloritoid assemblage is recognized (Cady and others, 1962; Albee, 1968; Lapp, 1986; Lapp and Stanley, 1986; O'Loughlin, 1986; O'Loughlin and Stanley, 1986). Metamorphism associated with the second generation fold structures (F3) is documented at a minimum at the biotite zone/greenschist facies east of the Lincoln massif and is not recognized west of the eastern part of the Lincoln massif. This event may be related to the Acadian orogeny or represent polymetamorphism during the Taconian orogeny. The existing isotopic age data does not allow differentiation between these interpretations.

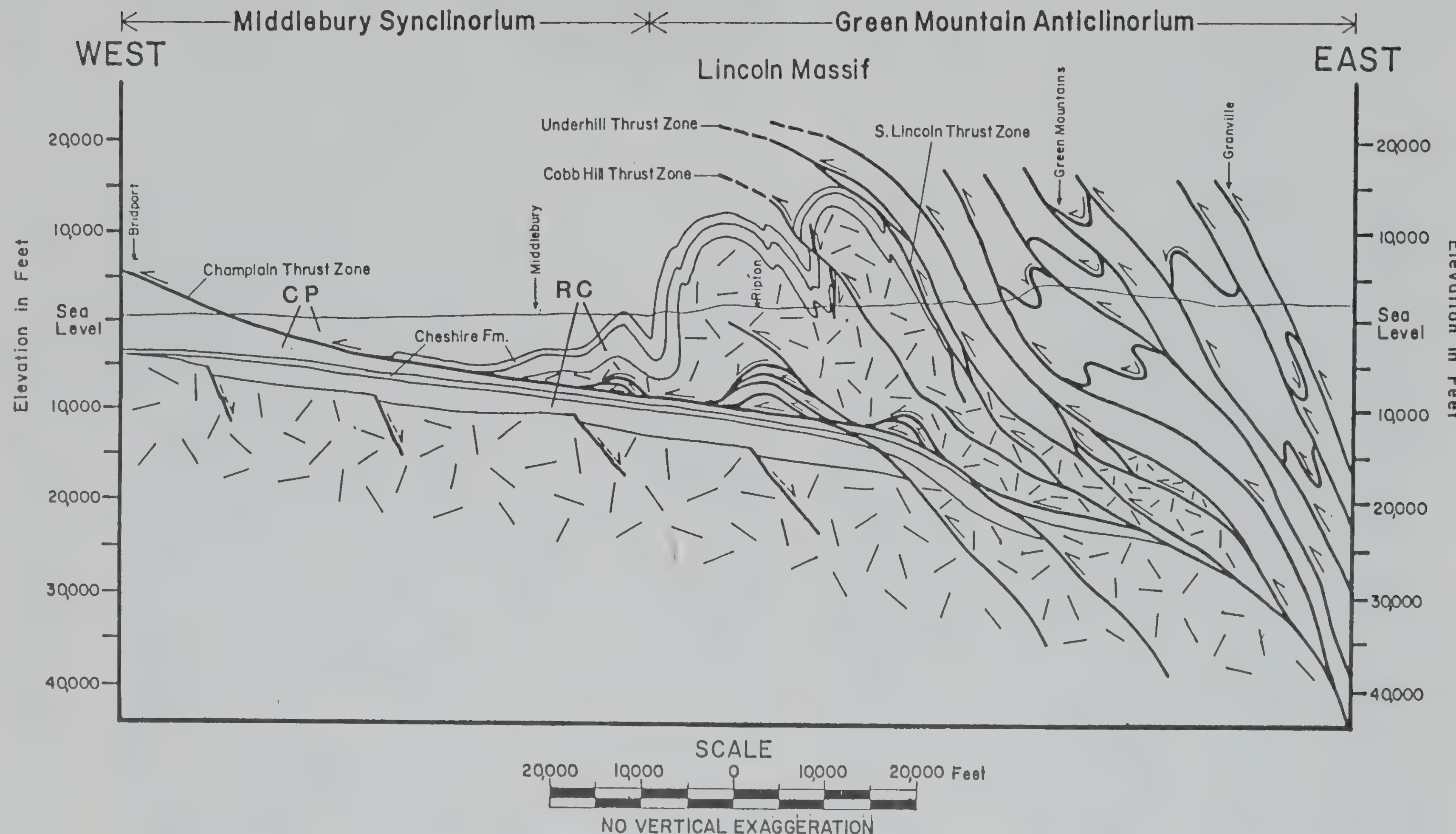
Regional Cross Section

The regional implications of the structural relations described in this study are illustrated in an interpretative east-west cross section located at latitude 43.00 N through the Lincoln massif (Fig. 3). The greatest assumption associated with drawing this section is in the location of the Champlain thrust zone. The displacement across this thrust zone is thought to be on the order of 45 to 50 km (Stanley and Ratcliffe, 1983; 1985) but may be as large as 80 km or more (Rowley, 1982). This zone was originally interpreted to root in basement rocks in the core of the Green Mountains (Doll and others, 1961, cross-sections A-A' and B-B'; Coney and others, 1972; Stanley and Ratcliffe, 1985).

Recent deep crustal seismic surveys across the Green Mountain massif in southern Vermont (Ando and others, 1983; 1984) indicate that a zone of prominent subhorizontal reflectors extends continuously from west to east through the Green Mountain massif (GMM) at depths between approximately 3,670 to 6,070 m (11,000 to 18,200 ft). The upper part of this zone has been correlated with the Champlain thrust fault at 3,670 m (11,000 ft) below the surface. The seismic data show that this zone roots beneath, and to the east of, the GMM. This implies that the exposed basement of the Green Mountain and Lincoln massifs are allochthonous and have been transported westward on the Champlain thrust. Furthermore, it has been suggested by Ando and others (1984) that part of the autochthonous platform sequence may be preserved below the Champlain thrust, extending beneath the exposed basement. These rocks would be represented by the 2400 m (7200 ft) interval of subhorizontal reflectors outlined above. This would imply that the top of the autochthonous basement occurs at a depth of at least 6,070 m (18,200 ft) below the surface of the GMM. Such an interpretation would be consistent with displacements across the Champlain thrust suggested by Stanley and Ratcliffe (1983; 1985). Alternatively, the absence of carbonate bank or rift-clastic rocks beneath the

Figure 3. Interpretive east-west cross-section along latitude 43.00 N through the Lincoln massif in central Vermont. This cross-section shows the Champlain thrust zone projected eastward beneath the Lincoln massif where it roots to the east of the exposed basement. Part of the autochthonous platform (CP) and rift-clastic (RC) sequence is shown to exist beneath the Champlain thrust zone. Anticlinal stacks or duplexes shown above the Champlain thrust zone are highly interpretive, but are believed to be an important mechanism in the development of the large-scale basement folds (Stanley and Ratcliffe, 1985). This cross-section shows the interpretation that the Lincoln massif represents a series of large-scale basement folds, which were progressively, internally imbricated from east to west and subsequently transported westward along the Champlain thrust zone as an allochthonous basement slice during the Taconian orogeny. Faults are shown as arrowed heavy lines. Lithic contacts shown by thin lines. The basement complex is represented by random dashes.

INTERPRETIVE CROSS-SECTION ALONG LATITUDE 44°00'N
THROUGH THE LINCOLN MASSIF — CENTRAL VERMONT



GMM would indicate a more limited displacement across the Champlain thrust and, in turn, imply that the Green Mountain and Lincoln massifs are parautochthonous. The cross-section illustrated in Figure 1 incorporates the former interpretation.

The western shelf sequence and underlying rift-clastics are attached to the Lincoln massif and are interpreted to have extended to the eastern limb of the Lincoln anticline. The Cheshire Formation is highlighted for clarity in that part of the cross-section. Evidence for this is presented by Thompson (1972; Chang and others, 1965) in the Plymouth, Vermont, area where dolostone horizons are thought to represent carbonate bank deposits correlative with the Lower Cambrian Dunham Formation of the Western Sequence. This would indicate that the western shelf sequence extended to the east of the GMM. In Figure 1, the Champlain thrust truncates the platform sequence showing a minimum displacement of approximately 20 km (13 mi), although the actual displacement across that zone is presently unknown.

The large scale folding of the Lincoln massif is thought to have been formed by fault-bend folds, or duplexes at depth, as well as by ductile flow associated with the development of cleavage in the basement rocks (Fig. 1; Stanley and Ratcliffe, 1985). These structures would have developed prior to, or synchronously with, the early displacement history of the Champlain thrust and would root along that zone.

The rocks east of the Lincoln massif were severely shortened by folding and subsequently intercalated by multiple generations of thrust faults (DelloRusso, 1986a; Lapp, 1986; Lapp and Stanley, 1986; O'Loughlin, 1986; O'Loughlin and Stanley, 1986; Stanley and others, 1985; 1986). The recognition of extensive thrust zones in that sequence results in a complex interpretation of the stratigraphy. For simplicity, lithic units are not distinguished and major thrust zones are labelled in Figure 1.

Importantly, the steep orientation of the structural fabric observed adjacent to the basement-cover boundary along the eastern margin of the Lincoln massif becomes distinctly more shallow along the crest of the Green Mountains (Lapp, 1986; Lapp and Stanley, 1986; O'Loughlin, 1986; O'Loughlin and Stanley, 1986) and steepens again to the east of the Green Mountain axis (Stanley and others, 1986). The flattening in the regional trend across the axis of the Green Mountains may be the result of a large-scale ramp zone at depth. If major thrust zones in the Eastern Sequence are projected westward, it becomes apparent that the Lincoln massif defines yet another basement ramp against which the Eastern Sequence must have abutted, resulting in additional shortening in the Eastern Sequence. This interpretation would explain the intensity of deformation and steep orientation of structures observed along the eastern limb of the Lincoln anticline and the structural contrast with western areas. It is evident that the Lincoln massif may represent a remnant piece of a larger basement ramp, located to the east, which was sheared off and transported westward along the Champlain thrust zone.

CONCLUSIONS

The information presented in this study relates to all aspects of the geology in central Vermont and affects regional geologic interpretations as well. As with any study of this scope, more questions are raised than are answered. It is hoped that the ideas, interpretations, and conclusions reached in this study will encourage further study. The list that follows outlines the most significant results of the study.

1. The Mt. Holly Complex of the eastern part of the Lincoln massif (ELM) is composed of both meta-igneous and metasedimentary rocks. In contrast, the western part of the Lincoln massif (WLM) contains a laterally continuous, mappable sequence of quartzite and paragneisses.

2. The two metasedimentary sequences in the ELM, blue quartzite (Ymhq)/ tourmaline-chloritoid schist (Ymht) and plagioclase augen gneiss/schistose quartzite, unconformably overlie an older basement of granitic gneiss. These two sequences were deformed into broad folds during the Grenvillian orogeny.

3. Relict Grenvillian metamorphism is poorly preserved but reached the epidote-amphibolite facies/garnet zone within the Mt. Holly Complex.

4. The granitic gneiss (Ymhg) of the ELM is intruded by mafic dikes (Zmd) of post-Grenvillian age. These dikes do not appear to transect the basal unconformity with the overlying Pinnacle Formation. They were intruded, therefore, prior to deposition of the cover sequence. It is likely that these dikes are associated with Late Proterozoic rifting of the basement.

5. A complete depositional sequence spanning rift to drift stages of rifting is preserved in the cover overlying the WLM and the western margin of the ELM. The Western Sequence is considered to be attached to the basement because Paleozoic faults are rare and of minor extent in the western part of the study area.

6. A series of amphibolite exposures within the Hoosac Formation defines a continuous N-S trending belt within the Eastern Sequence. This indicates stratigraphic continuity between the Hoosac and Underhill Formations at the northern end of the ELM, rather than a pinch-out of the Hoosac Formation as previously proposed.

7. A bedding-plane schistosity (S1) predates the development of the dominant schistosity (S2) and major folds in the area (F2). F2/S2 deformation developed during epidote-amphibolite facies metamorphism (peak-M1) in the east and during biotite zone/greenschist facies metamorphism in the west.

8. Crenulations and east-facing, open folds (S3/F3) deform the S2 schistosity east of the ELM and are associated with retrogressive metamorphism at biotite zone conditions (post-peak M1 or M2?).

9. A major west-dipping, high-angle shear zone, displaying east-side-down relative displacement, is associated with sheared-out parasitic folds along

the western limb of the Lincoln anticline.

10. An extensive north-south trending zone of mylonites within the Mt. Holly Complex comprises the Cobb Hill thrust zone. This zone defines the western limit of major east-over-west thrust zones within the Lincoln massif.

11. The South Lincoln and Underhill thrust zones represent major, regionally continuous zones of east-over-west displacement which incorporate basement slivers along the eastern limb of the Lincoln anticline. The Eastern Sequence west of the South Lincoln thrust zone is, therefore, allochthonous with respect to the ELM.

12. The Lincoln massif represents a series of large-scale basement folds that were progressively imbricated from east to west. The massif was subsequently transported westward along the Champlain thrust as an allochthonous basement slice, during the Taconian orogeny.

ACKNOWLEDGEMENTS

We gratefully acknowledge Barry L. Doolan, Murray Journey, and Sharon O'Loughlin for their critical reviews and comments that greatly improved the final manuscript. This research was made possible by funding from the following sources: the U.S. Forest Service (U.S. Department of Agriculture contract 524526, awarded to Rolfe Stanley for mapping in the Green Mountain National Forest), the National Science Foundation (grant EAR 8516879, awarded to Rolfe Stanley), and the U.S. Department of Energy (grant DE-FG02-81WM46642, awarded to the Vermont Geological Survey). Any opinions, findings, conclusions, or recommendations expressed herein are those of the authors and do not necessarily reflect the views of the Department of Energy.

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Bedrock Geology of the Northern Part of the Lincoln Massif, Central Vermont

Compiled By

Vincent DelloRusso and Rolfe S. Stanley

PLATE 1

LITHOLOGIC DESCRIPTIONS

Western Sequence

Cheshire Formation

€c — **Massive Quartzite** White, grey to pink, massive, fine-to medium-grained quartzite and argillaceous quartzite containing minor phyllite. Discontinuous, mottled, white, rippled quartzite beds from 2.5 to 20 cm. thick occur locally.

Fairfield Pond Formation

€Zfp — **Chlorite Phyllite** Grey, grey-green, light-grey to light-brown weathering, fine-grained quartz-sericite-chlorite phyllite. Biotite is characteristically rare. Fine, light-grey laminations common. Magnetite common near base of unit.

Pinnacle Formation

€Zpcl — **Chlorite-Magnetite Schist** Light-green, locally rusty, silvery-grey weathering, fine-grained quartz-plagioclase-muscovite-magnetite-chlorite schist. Abundant euhedral magnetite porphyroblasts up to 7mm. in diameter are diagnostic. Locally interbedded with chlorite-rich, pebbly metagreywacke. Associated with discontinuous pods of vein quartz and dolomitic interbeds.

€Zpfd — **Forestdale Dolarenite** Light-brown, dark-brown to black weathering, sandy dolarenite/calclithite. Typically massively bedded containing thin chlorite schist interbeds. Subangular grains of quartz and feldspar up to 7mm. in diameter are common. Beds up to 0.5m. thick exhibit normal grading of clastic grains. Outcrop form ridges up to 8m. high.

€Zpbg — **Biotite Metagreywacke** Grey to dark-grey, mottled brown weathering, medium-to coarse-grained, massively bedded, quartz-feldspar-muscovite-chlorite-biotite metawacke. Locally containing conglomeratic beds up to 1m. thick with well-rounded quartz and gneiss pebbles up to 5cm. in length. Mottled weathering due to minor carbonate content. Characteristic "Pinnacle" lithology.

€Zpm — **Muscovite Metagreywacke** Silvery, light-grey to grey-green, rusty-brown weathering, fine-grained, massive, quartz-feldspar-muscovite metawacke containing minor amounts of chlorite, magnetite, epidote, and tourmaline. Biotite is characteristically rare. Rock fragments are conspicuously absent. Quarter inch laminations are locally well-developed.

€Zpbc — **Conglomerate** Poorly sorted, matrix-supported, quartz and gneiss

EXPLANATION

Observed bedrock exposures.

Lithic contacts. Everywhere solid. Accuracy of location indicated by proximity to bedrock exposures.

Axial traces of major folds. Dashed where highly interpretive.

Overturned syncline.

Overturned anticline.

Thrust fault/Shear zone. Dashed where interpretive; solid where fault evidence observed. Teeth on upper plate. Tick mark indicates dip orientation. Strike-slip component indicated by half-arrows.

High angle fault/Shear zone. Dashed where interpretive; solid where fault evidence observed. Sense of relative motion across fault designated by U (upthrown block) and D (downthrown block) symbols. Tick marks indicate dip orientation.

Fault or shear zone for which sense of motion is unknown. Dashed where interpretive; solid where fault evidence observed. Tick mark indicates dip orientation.

Planar Features

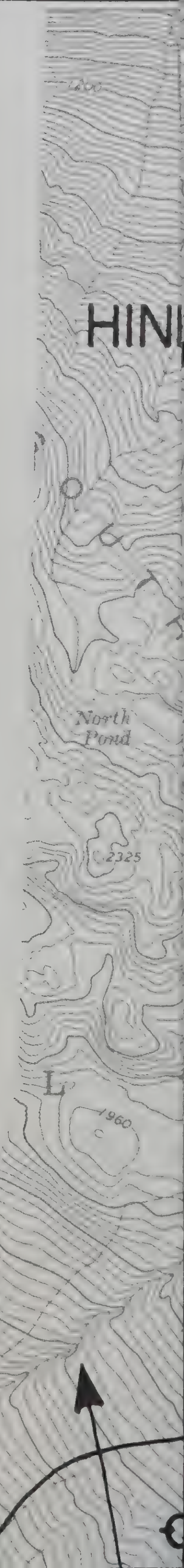
Strike and dip of bedding in metasedimentary rocks, and compositional layering in Ymhml.

Strike and dip of Grenvillian schistosity in gneisses, defined by flattened quartz grains and aligned micas.

Strike and dip of bed-parallel schistosity (S1). Defines folds to which the dominant schistosity is axial planar.

Strike and dip of the dominant schistosity (S2). Parallel to the axial surface of the folded bed-parallel schistosity in the western cover rocks. Deformed into tight to isoclinal minor folds and crenulations within the eastern cover rocks.

Strike and dip of the axial surface to folds that deform the bed-parallel schistosity. The dominant schistosity is observed parallel to this surface. Defines F2 folds.



Bedrock Geology of the Northern Part of the Lincoln Massif, Central Vermont

Compiled By
Vincent DelloRusso and Rolfe S. Stanley

PLATE 1

LITHOLOGIC DESCRIPTIONS

Western Sequence

Cheshire Formation

€c — Massive Quartzite. White, grey to pink, massive, fine-to medium-

Fairfield Pond Formation

€Zfp — Chlorite Phyllite

Pinnacle Formation

€Zpcl — Chlorite-Magnetite Schist

€Zpfd — Forestdale Dolarenite

€Zpbg — Biotite Metagreywacke

€Zpm — Muscovite Metagreywacke

€Zpbc — Conglomerate

Eastern Sequence

Underhill Formation

€Zu — Undifferentiated Schists. Dominantly silvery, rusty weathering, ... grained garnetiferous plagioclase-quartz-muscovite schist

€Zuql — Quartz-laminated schist

Hoosac Formation

€Zhg — Schistose Metagreywacke

€Zhms — Mica schist

€Zhbg — Biotite Metagreywacke

€Zhbc — Conglomerate

Mount Holly Complex

Zmd

Ymhg

Ymht

Ymhaq

Ymhag

Ymhml

Ymhgb

Ymha

Ymhbg

Ymhgq

Ymhga

Ymhg

EXPLANATION

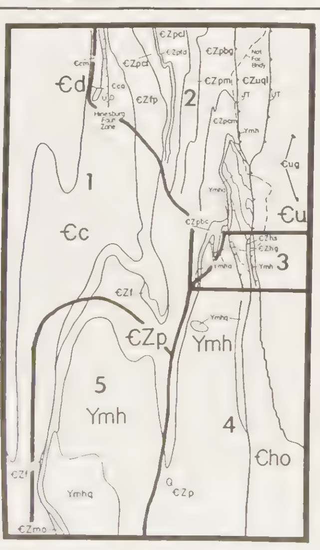


Figure 1. Generalized Geologic map of the northern part of the Lincoln massif outlining map areas of the following workers whose contributions to this compilation map (Plate 1) are gratefully acknowledged: 1) C. Doll, et. al. (1961). 2) & 3) P. Tauvers (1982). 3) & 4) V. DelloRusso (1986). 5) J. Prahl (1985).

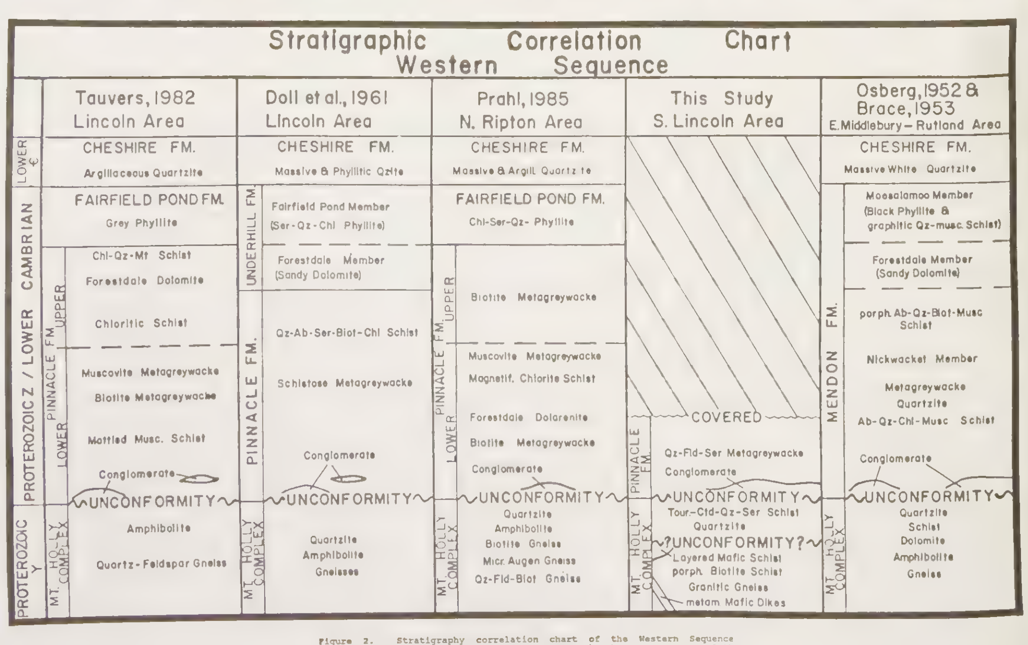


Figure 2. Stratigraphy correlation chart of the Western Sequence Unconformably overlying the Mt. Holly complex in the western part of the Lincoln massif are the rift-clastic rocks of the Pinnacle Formation, overlain by rift-drift transition rocks of the Fairfield Pond and Cheshire Formations. Note the varied stratigraphic position of the Forestdale dolarenite, which is interpreted as being deposited on local basement topographic highs (horsts) within the proto-Atlantic rift basin during sedimentation of the Pinnacle wackes (Tauvers, 1982). Within the Mt. Holly complex, a sequence of tourmaline-chloritoid schist and quartzite overlies the granitic gneiss of the western part of the ELM within the Mt. Holly complex, but predates deposition of the Pinnacle Formation. Detailed stratigraphic relations are described in the text.

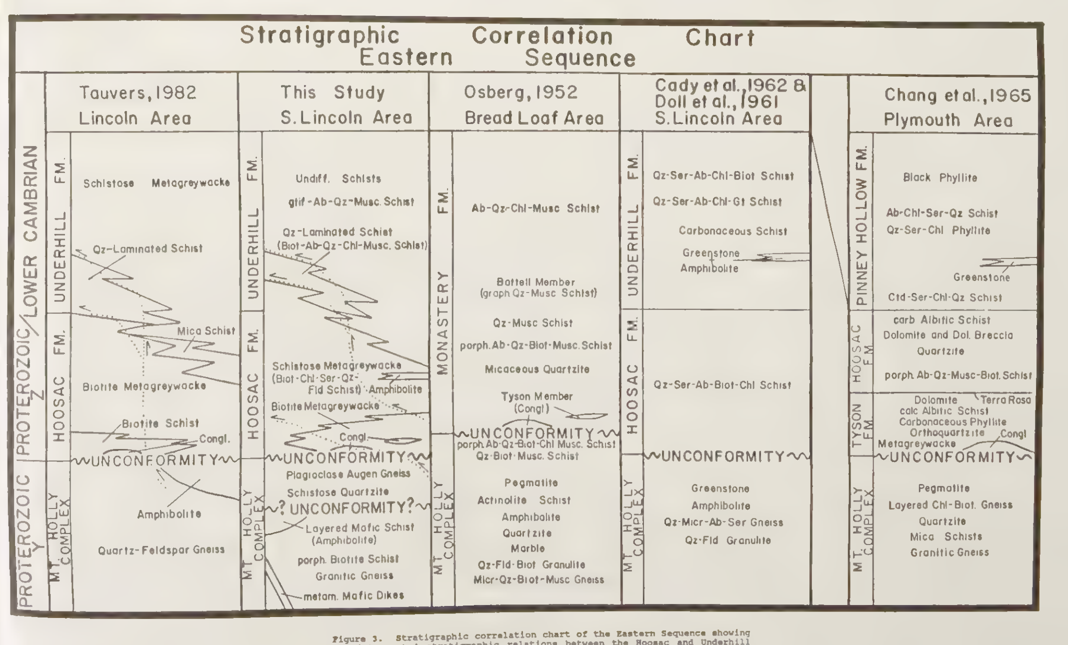


Figure 3. Stratigraphic correlation chart of the Eastern Sequence showing the interpreted stratigraphic relations between the Hoosac and Underhill Formations. The relative positions of major thrust faults are shown as arrowed, dotted lines. Note that the basal part of the Hoosac Formation in the Lincoln-So. Lincoln area is correlative to part of the Tyson Formation of the Plymouth area. Within the Mt. Holly complex, a sequence of plagioclase augen gneiss and quartzite overlies the granitic gneiss of the eastern part of the ELM. Detailed structural and stratigraphic relations are described in the text.

Bedrock Geology of the Eastern Massif, Central Vermont

PLA

Vincent DelloRusso

SCALE

.5 .25 0 .5

MILES

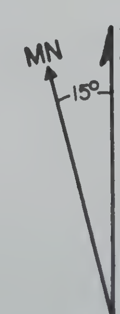
2000 1000 0 2000

FEET

.5 .25 0 .5

KILOMETERS

Contour interval 20 feet



LITHOLOGIC DESCRIPTIONS

Western Sequence

Pinnacle Formation

€Zpm **Schistose Metagreywacke** Light-grey to rusty-brown weathering, plagioclase-quartz-sericite schist/feldspathic metawacke containing minor amounts of microcline, perthite, biotite, chlorite, epidote, carbonate, tourmaline, zircon, and opaques. Relict detrital grains of feldspar and blue quartz are common.

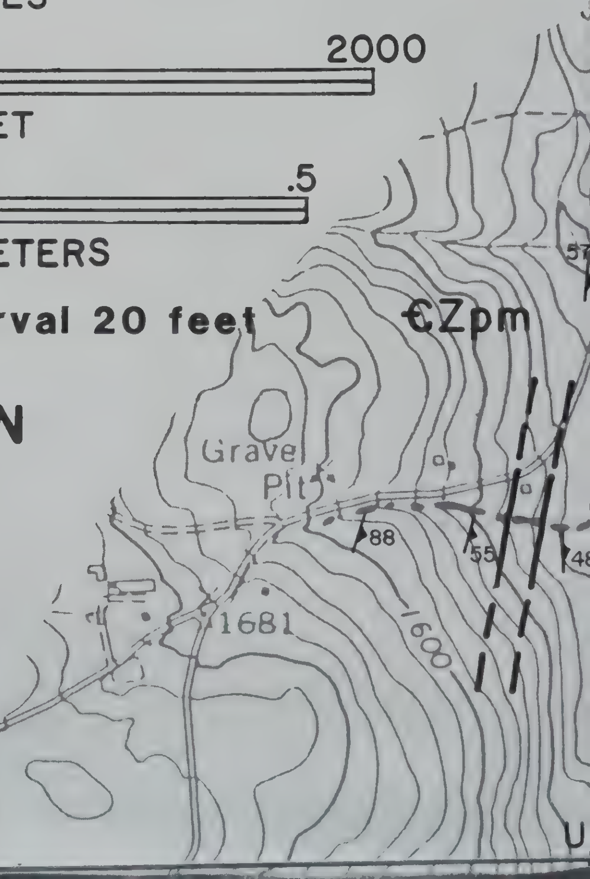
€Zpbc **Conglomerate** Massive, matrix-supported, quartz cobble conglomerate containing rounding quartz clasts up to 15cm in diameter, and gneiss (Ymhg) boulders up to 55cm in diameter within a matrix compositionally similar to CZpm.

Eastern Sequence

Underhill Formation

€Zu **Undifferentiated Schists** Dominantly silvery, rusty weathering, medium-grained, garnetiferous plagioclase-quartz-muscovite schist containing minor amounts of biotite, chlorite, epidote, and opaques. Locally quartz-feldspar rich metawacke resembling CZhg.

€Zuql **Quartz Laminated Schist** Light-grey to rusty weathering, medium-grained, garnetiferous biotite-plagioclase-quartz-chlorite-muscovite schist containing minor amounts of magnetite. Displays characteristic 3-4mm thick quartz laminations, separated by mica-rich segregations.



Bedrock Geology of the Eastern Lincoln Massif, Central Vermont

PLATE 2

Vincent DelloRusso

LITHOLOGIC DESCRIPTIONS

Western Sequence

Pinnacle Formation

€Zpm **Schistose Metagreywacke** Light-grey to rusty-brown weathering, plagioclase-quartz-sericite schist/feldspathic metawacke containing minor amounts of microcline, perthite, biotite, chlorite, epidote, carbonate, tourmaline, zircon, and opaques. Relict detrital grains of feldspar and blue quartz are common.

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Eastern Sequence

Underhill Formation

€Zu **Undifferentiated Schists** Dominantly silvery, rusty weathering, medium grained, garnetiferous plagioclase-quartz-muscovite schist containing minor amounts of biotite, chlorite, epidote, and opaques. Locally quartz feldspar rich metawacke resembling CZhg.

€Zuql **Quartz Laminated Schist** Light-grey to rusty weathering, medium-grained, garnetiferous biotite-plagioclase-quartz-chlorite-muscovite schist containing minor amounts of magnetite. Displays characteristic 3-4mm thick quartz laminations, separated by mica-rich segregations.

Hoosac Formation

€Zhg **Schistose Metagreywacke** Light-grey to white, locally rusty weathering, fine- to medium-grained sericite-quartz-plagioclase schist/feldspathic metawacke containing minor amounts of garnet, biotite, chlorite, epidote, and opaques. Primary clasts are rare.

€Zhms **Mafic Schist** Light-grey and dark-green to black, garnet-biotite-chlorite-quartz-plagioclase amphibolite containing minor amounts of epidote, sphene, magnetite, and zircon. The rock appears pin-striped due to quartz-plagioclase/amphibole segregation. Interlayered with feldspathic metawacke, CZhg.

€Zhbg **Biotite Metagreywacke** Light-grey to dark-green-brown fine- to medium-grained, chlorite-biotite-quartz-plagioclase schist/feldspathic biotite metawacke containing minor amounts of sericite, epidote, carbonate, magnetite, and zircon. Locally magnetite-rich. Interlayered with lenses of conglomerate, CZhbc.

€Zhbc **Conglomerate** Quartz, feldspar, and gneiss cobble to pebble conglomerate. Matrix is compositionally similar to CZhbg. At location D-19, stretched quartz and gneiss cobbles are up to 30cm long, in a chlorite-rich schistose matrix. At location X-20, clasts are primarily composed of plagioclase up to 2cm in diameter, in a biotite-rich schistose matrix.

Mount Holly Complex

Zmd **Mafic Dikes** Dark-green-brown weathering, fine- to medium-grained, sericite-epidote-biotite-chlorite-plagioclase schist containing minor amounts of carbonate, sphene, and opaques. Commonly altered to biotite-rich schist along contact zones. Large, relict, euhedral, white plagioclase phenocrysts up to 2cm in length are characteristic of these intrusives. Xenoliths of Ymhg are abundant at location S-10.

Ymhq **Blue Quartzite** Clean, massive, light-blue to light-grey weathering quartzite containing minor amounts of sericite, biotite, chlorite, plagioclase, rutile, and opaques. Displays broadly-folded beds from 0.5m to 1m thick. Interbedded with Ymht.

Ymht **Tourmaline-Chloritoid Schist** Rusty-weathering, fine-grained, quartz-chloritoid-tourmaline-sericite schist containing minor amounts of chlorite, graphite, and zircon. Chloritoid occurs as large euhedral porphyroblasts and as radiating, blade-like grain aggregates forming spherical "suns". Graphite is locally abundant in seams up to 5cm thick. This unit is bounded by thick massive beds of Ymhq.

Ymhaq **Schistose Quartzite** Massive, blue-green weathering plagioclase-chlorite-sericite schistose quartzite containing minor amounts of epidote and opaques. Interbedded with fine-grained augen gneiss. Exposures form steep slopes along the ridge southeast of Cobb Hill.

Ymhag **Plagioclase Augen Gneiss** Light-grey to white, and dark-green-brown weathering, coarse-grained sericite-chlorite-quartz-biotite-plagioclase augen gneiss containing minor amounts of epidote, sphene, rutile, zircon, and opaques. Large white plagioclase porphyroclasts up to 4cm in diameter are characteristic, and occur within a biotite-rich schistose matrix. Locally containing beds of Ymhaq.

Ymhml **Layered Mafic Schist** Dark green to black, medium- to coarse-grained garnet-plagioclase amphibolite retrograded to epidote-chlorite-quartz-biotite-actinolite-plagioclase schist. Contains minor amounts of carbonate and sphene. Thin, white, sigmoidal plagioclase-rich layers are characteristic of this unit. Contains xenoliths of Ymhg.

Ymhgb **Biotite Schist** Dark-brown to black weathering, fine-grained quartz-sericite-biotite schist containing minor amounts of albite, chlorite, epidote, carbonate, and opaques. Coarse-grained, randomly oriented biotite porphyroblasts are typically abundant. Occurs as narrow, foliated zones within Ymhg.

Ymhg **Granitic Gneiss** Light-grey to white weathering, relatively massive, medium-grained muscovite-perthite-microcline-quartz-plagioclase gneiss containing minor amounts of garnet, biotite, chlorite, epidote, carbonate, zircon, and magnetite.

THE ORDER OF MAP UNITS ABOVE DOES NOT NECESSARILY REFLECT THEIR RELATIVE STRATIGRAPHIC POSITION

EXPLANATION

Observed bedrock exposures.

Lithic contacts. Everywhere solid. Accuracy of location indicated by proximity to bedrock exposures.

Axial traces of major folds. Dashed where highly interpretive.

Overturned syncline.

Overturned anticline.

Thrust fault/Shear zone. Dashed where interpretive; solid where fault evidence observed. Teeth on upper plate. Tick mark indicates dip orientation. Strike-slip component indicated by half-arrows.

High angle fault/Shear zone. Dashed where interpretive; solid where fault evidence observed. Sense of relative motion across fault designated by U (upthrown block) and D (downthrown block) symbols. Tick marks indicate dip orientation.

Fault or shear zone for which sense of motion is unknown. Dashed where interpretive; solid where fault evidence observed. Tick mark indicates dip orientation.

Brittle normal fault displaying minor displacement. Fault crosscuts all other rock fabric. Hachures are on the hanging wall block, and indicate the dip orientation. Arrow defines slickenside lineation.

Planar Features

Strike and dip of bedding in metasedimentary rocks, and compositional layering in Ymhml.

Strike and dip of Grenvillian schistosity in gneisses, defined by flattened quartz grains and aligned micas

Strike and dip of bed-parallel schistosity (S1). Defines folds to which the dominant schistosity is axial planar.

Strike and dip of the dominant schistosity (S2). Parallel to the axial surface of the folded bed-parallel schistosity in the western cover rocks. Deformed into tight to isoclinal minor folds and crenulations within the eastern cover rocks.

Strike and dip of the axial surface to folds that deform the bed-parallel schistosity. The dominant schistosity is observed parallel to this surface. Defines F2 folds.

Strike and dip of the axial surface to minor folds and crenulations that deform the dominant schistosity. There is no macroscopic cleavage observed parallel to the axial surface of these folds. Defines F3 folds.

Strike and dip of prominent joints.

Linear Features

Trend and plunge of fold axes of minor F2 and F3 folds. Arrow indicates plunge direction. Sense of rotation shown for asymmetrical folds as viewed in down-plunge direction.

Clockwise rotation sense.

Counterclockwise rotation sense.

Indeterminate rotation sense or symmetrical fold.

Trend and plunge of the intersection lineation of bed-parallel schistosity (S1) and the dominant schistosity (S2).

Trend and plunge of the axis of crenulations.

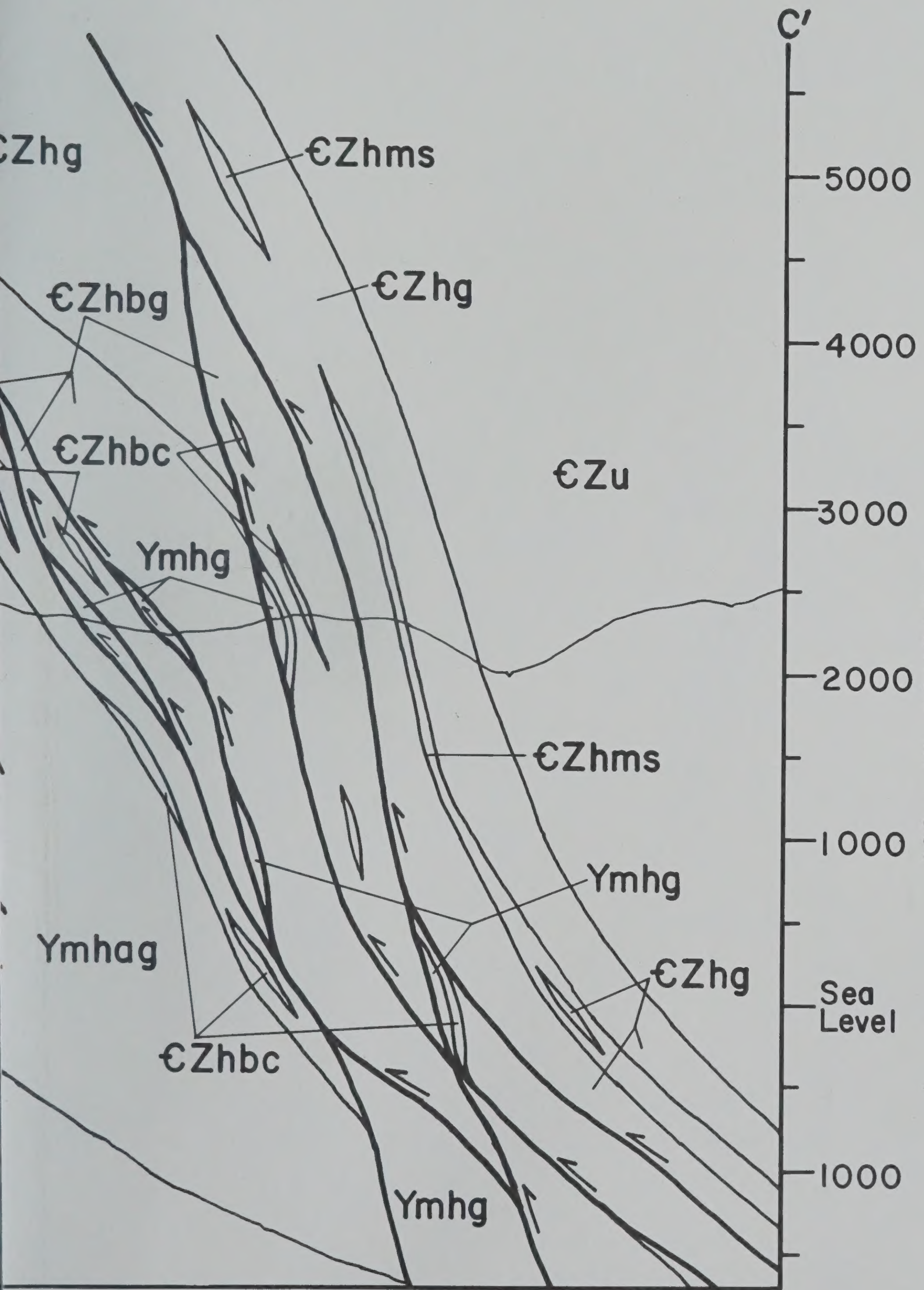
Trend and plunge of mineral lineations, including quartz rods, and stretching lineations on fault surfaces.

A A' Location of cross-sections.

Special Bulletin No. 8, 1986
DelloRusso and Stanley
Vermont Geological Survey
Charles A. Ratte, State Geologist

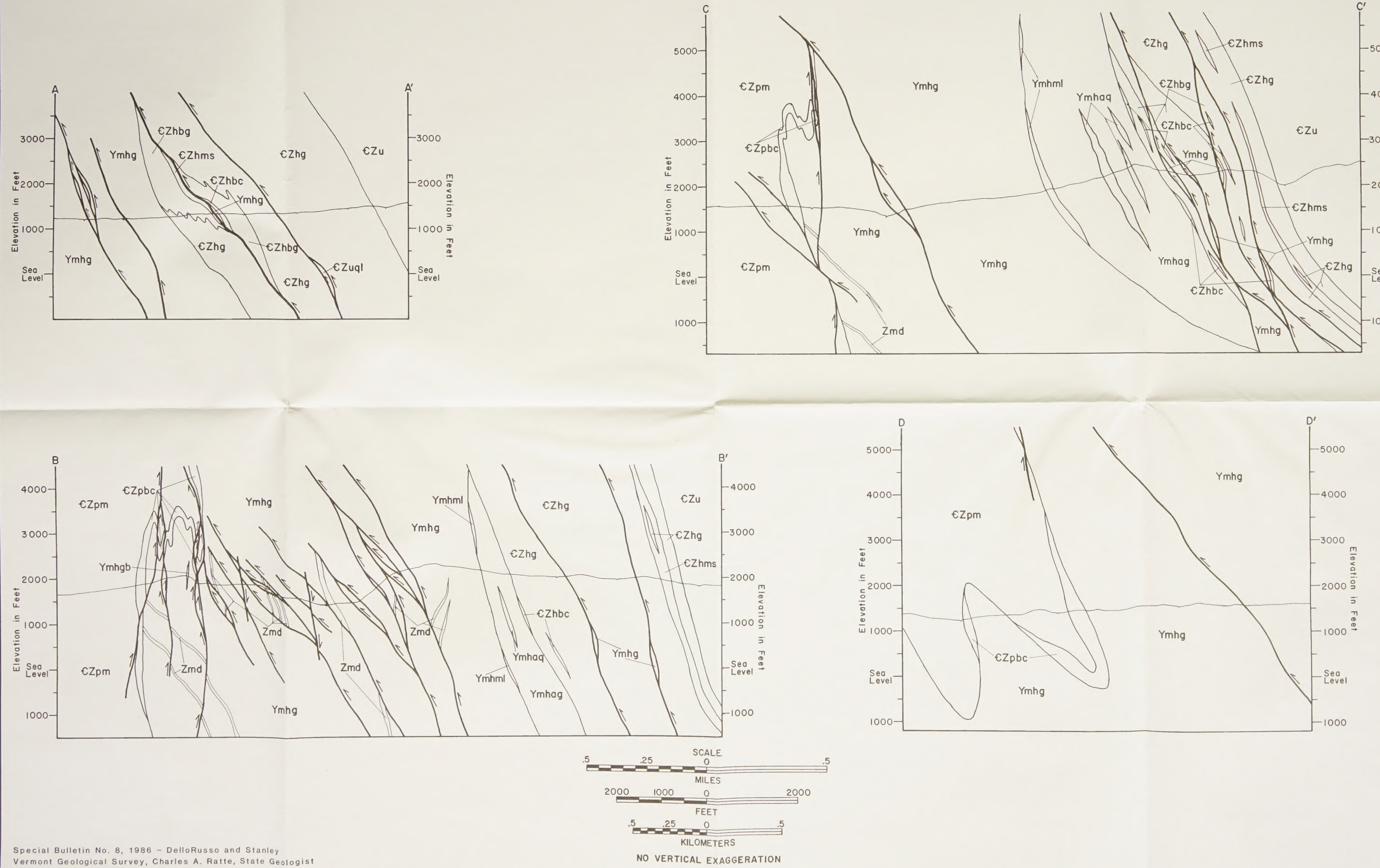
Figure 1. Generalized stratigraphic correlation chart of the eastern Lincoln massif. This diagram shows the interpretive stratigraphic position of map units within the study area. The Hoosac, Underhill and Pinnacle Formations represent the same stratigraphic level, although not strictly illustrated as such in this diagram. The relative position of major fault zones is shown. True dip angles and displacements are not represented. Note that the Hoosac and Underhill Formations are interpreted as an eastern facies equivalent of the Pinnacle Formation. Mafic dikes (Zmd) are represented as the same age as the basement-cover unconformity, because they do not truncate that surface, but truncate the Grenvillian fabric in the basement gneiss.

PLATE 3


C′
€Zhg
€Zhms
5000
€Zhbg
€Zhg
4000
€Zhbc
€Zu
3000
Ymhg
Elevation in Feet
2000
€Zhms
1000
Ymhg
Ymhag
€Zhg
Sea Level
€Zhbc
1000
Ymhg

Cross Sections

PLATE 3



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Vermont Geological Survey, Charles A. Ratte, State Geologist

PLATE 4

3 - Folded Mylonitic Schistosity

ty is observed in the same exposure as
escribed viewed looking east. The
is well-developed and defined by thin
Sm schistosity is folded by 1 to 2 cm
common north-over-south asymmetry when
these folds are sub-parallel to the
Fig. 16). This is the only exposure
preserves the folded Sm schistosity,
able, marking the transition from a
ase Two) to shear band development

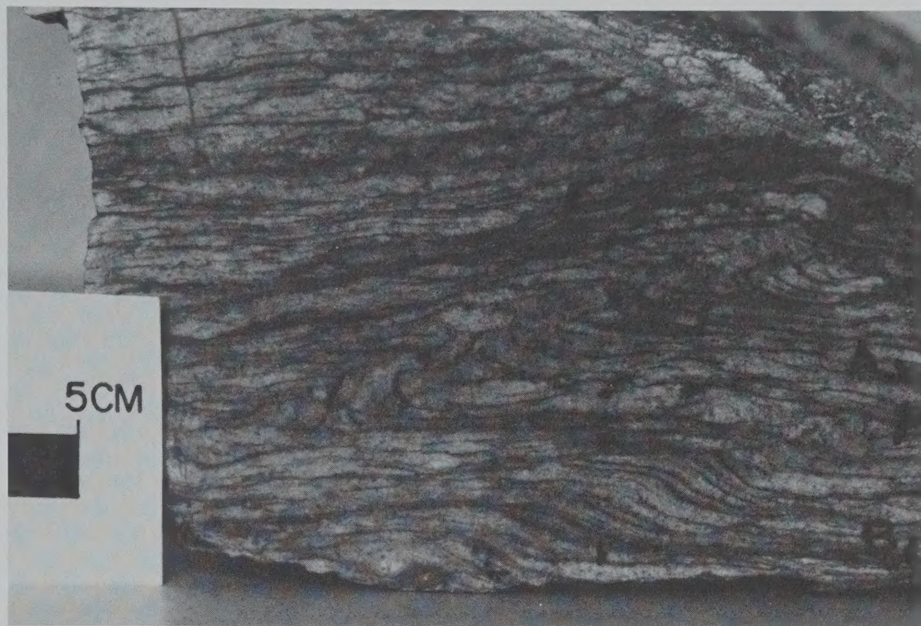


Figure 9. In this hand specimen (above), the folded mylonitic schistosity, Sm, is viewed looking east in the down-plunge direction. Note the few thin micaceous (dark) bands that transect these folds (center of photo). These thin bands develop into a spaced, shear band foliation (Phase Four) which forms as a result of shearing out minor folds along their attenuated limbs. In the lower part of the photo, Sm is only weakly folded.

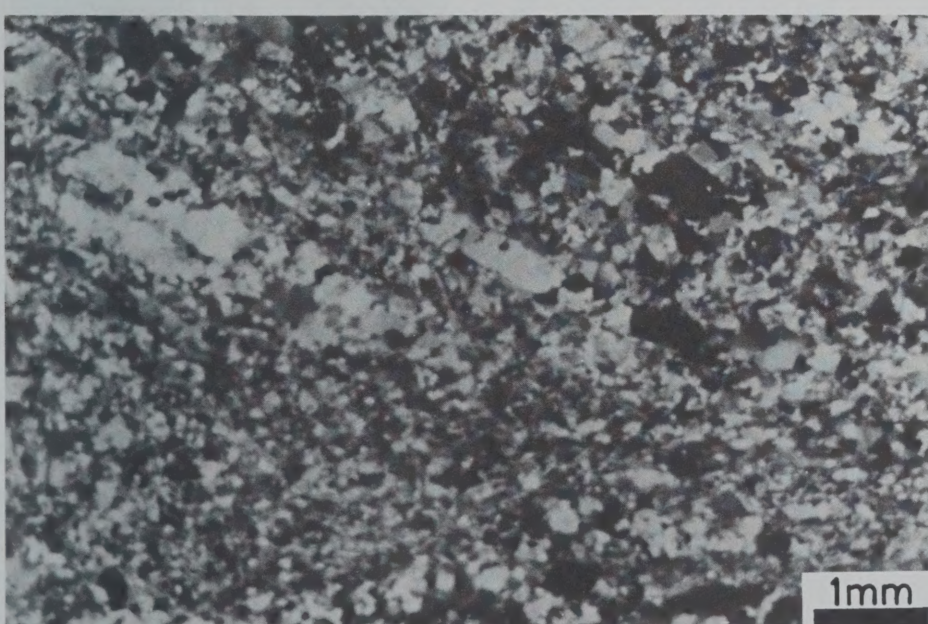


Figure 10. The fine-grained, strongly recrystallized quartz-feldspar-rich mylonitic fabric of the rock is illustrated in the photomicrograph above, viewed parallel to the mineral lineation in the rock. The Sm schistosity is outlined by elongate quartz ribbons (upper left to lower right). Note the bending of these ribbons (center). Sericite is common, but typically not abundant, suggesting that chemical alteration of feldspar was not extensive.

Shear Bands and Fragmented Mylonite

Evolution of the Cobb Hill Thrust Zone

PLATE 4

Introduction

The evolution of fault zone rocks preserved within the Cobb Hill thrust zone is best displayed within a series of three closely-spaced outcrops at location V13 (Plate 2). At this locality, four distinct phases of fabric development are observed: 1) gneiss fragmentation and development of an anastamosing mylonitic foliation, 2) pervasive mylonitic foliation, 3) folded mylonitic foliation, 4) fragmentation of the folded mylonitic foliation as a result of shear-band development. The intrusion of silica-rich, hydrous fluids during the later stages of the deformation is thought to result in extensive alteration of the gneiss, producing sericite phyllonite and vein quartz zones.

Phase 1 - Fragmentation of Gneiss

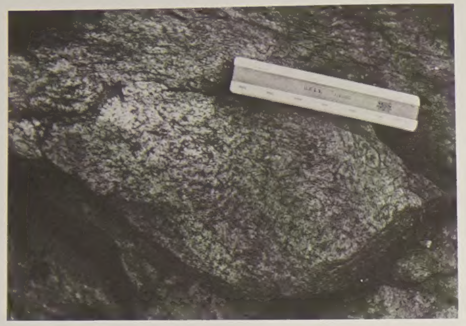



Figure 1 and 2. Outcrop photos of Phase One fabric at Loc. V13 (Pl.2) viewed looking east. The phase one fabric is characterized by large, angular to rounded clasts of coarse-grained gneiss up to 30 cm in diameter. The clasts record only weak Paleozoic deformation compared to the well-foliated, finer grained matrix that is composed of strongly deformed and altered gneiss. Major divisions on scale equal one inch (Fig. 1). Penny for scale (Fig. 2).

Figure 3. Outcrop photo of the Phase One fabric at Loc. V13 (Pl.2) viewed looking east. Fragments of gneiss commonly display the relict Grenvillian schistosity, which is discordant to the surrounding Paleozoic schistosity in the matrix. The matrix foliation anastamoses around the gneiss clasts and begins to overprint the clasts as they are progressively reduced in size. The angular and weakly deformed character of these clasts suggest that brittle fracturing of the gneiss is the dominant deformation mechanism during the initial stage of grain size reduction in the fault zone. The lack of sericite in the matrix indicates that chemical alteration of feldspar is not important during this stage of the deformation. In the rock matrix, dynamic recrystallization is the dominant deformation mechanism resulting in its fine grain size. In thin section (not shown), abundant coarse-grained porphyroclasts of feldspar are fragmented and sericitized along fractures. In the matrix surrounding the porphyroclasts, fine-grained, recrystallized feldspar and quartz are abundant. It appears, therefore, that recrystallization of feldspar is important only after the original grains have been extensively reduced in size by fracturing. Both coarse- and fine-grained quartz display features that are typical of dynamic recrystallization. Major divisions on scale equal one inch.

Phase 2 - Pervasive Mylonitic Schistosity

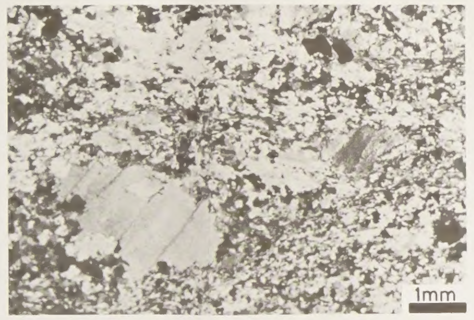
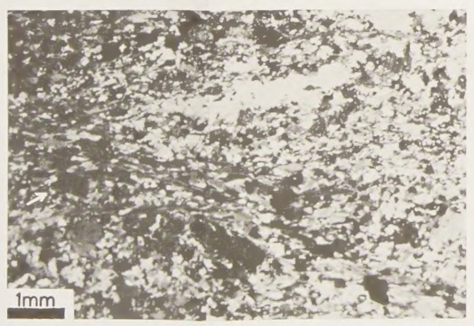


Figure 4. Hand specimen photo of a gneiss clast in the Phase Two fabric. The Phase Two fabric is displayed in an exposure approximately 32 m (100 ft.) north of the previous exposure. In this outcrop, clasts of coarse-grained granitic gneiss are abundant within a strongly recrystallized, mylonitic matrix. The clasts are strongly flattened into the mylonitic schistosity (Sm) and are elongate parallel to the stretching lineation in the rock. In contrast to the Phase One fabric, the mylonitic schistosity is very planar, and pervades the gneiss clasts. The clasts are lens-shaped and much smaller than those observed in Phase One. The fabric within the clasts (shown above) is parallel to and continuous with the matrix schistosity. It is evident, that the Grenvillian fabric, which is well-preserved in the Phase One fabric, is now obliterated from the gneiss fragments.



Figures 5 and 6. Oriented photomicrographs from the sample in Figure 4 illustrate the strongly recrystallized nature of quartz, and the brittle character of large feldspar porphyroclasts. The serrate nature of the porphyroclast grain boundary in Figure 5 and the abundance of fine-grained feldspar in the matrix indicates that dynamic recrystallization dominates once the grain size has been extensively reduced by fracturing. The imbrication of elongate quartz ribbons, and sheared feldspar porphyroclasts in figure 6 (viewed as if looking south with respect to outcrop) indicate a right-lateral shear sense in the photo, and an east-over-west shear sense in outcrop.

Figure 7. Hand specimen photo of the Phase Two fabric viewed as if looking north at the outcrop. As the gneiss clast size is reduced, and the mylonitic schistosity becomes pervasive. These clasts develop an asymmetric shape due to the dominant east-over-west simple shear component of deformation in the zone. An east-over-west shear sense is indicated by the asymmetric shape of the relict gneiss clasts, and the C-S fabric in the matrix. This shear sense is consistent with east-over-west sense of shear in the outcrop.

Summary of the Fault Zone Fabric Evolution

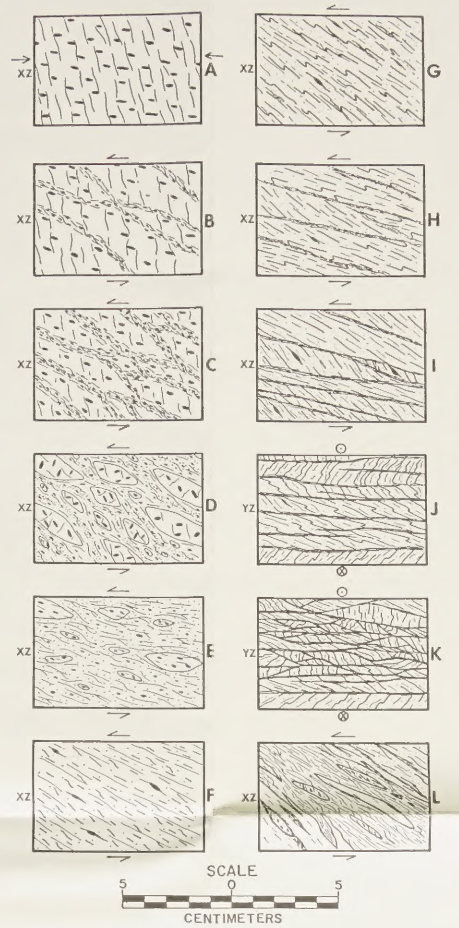
Phase 1

The initial stage of deformation within the Cobb Hill thrust zone resulted in fragmentation of the coarse-grained gneiss as evidenced by the numerous relict clasts of gneiss observed along the zone (Figure 15, A to D). Importantly, the matrix is of a similar composition to the gneiss, but the grain size is much finer. It is obvious that extensive grain size reduction accompanied the early stage of deformation. It is not clear, however, as to what extent brittle fracturing was important in this process. The recrystallized nature of quartz and feldspar obscures evidence of earlier fracturing. The abundance of fine-grained, recrystallized feldspar in the matrix suggests that crystal-plastic processes dominated subsequent to any earlier brittle deformation. It is also apparent that this process resulted in an increase of sericite in the rock matrix. Retrogression of feldspar, therefore, must have accompanied the deformation. In effect, the mica-forming reaction may have softened the rock, allowing it to deform ductilely. The lack of a strong fabric within the gneiss clasts, in contrast to the strongly foliated, mylonitic matrix emphasizes their resistance to ductile deformation.

After the initial fragmentation of the gneiss, the gneiss blocks behaved passively, as strain was accomodated by matrix deformation. The abundance of gneiss clasts results in an anastamosing geometry to the matrix foliation. The fine-grained nature of the matrix and greater mica content would allow crystal-plastic deformation mechanisms to dominate. The asymmetric geometry of C-S fabrics, and clast/porphyroclast tails preserve a consistent east-over-west shear sense, indicative of pervasive simple shear throughout the zone.

Phase 2

With increasing strain in the zone, dynamic recrystallization of quartz and feldspar further reduced the size of matrix grains and clasts until a pervasive, planar mylonitic schistosity was developed (Fig. 15, E and F). Segregation of quartz and feldspar is commonly observed as alternating light and dark layers less than 1 mm thick. Elongate quartz rods define the direction of elongation in the shear zone. The lack of relict clasts or porphyroclasts associated with this fabric suggests strain hardening of the fine-grained matrix, resulting in increased strain-induced recrystallization of the coarse-grained clasts and porphyroclasts. The size reduction and obliteration of clasts and porphyroclasts result from continued recrystallization at the clast/matrix contact, due to the inability of the fine-grained matrix to accomodate the imposed strain.



Phase 3

The pervasive mylonitic fabric was subsequently deformed into 1 to 2 cm scale asymmetric folds, illustrating the instability of the preexisting mylonitic fabric (Fig. 15G). The mylonitic fabric may have become unstable due to: 1) a decrease in temperature, 2) an increase in strain rate, 3) an effective work hardening of the rock, or 4) a rotation of the foliation with respect to the imposed strain field. Folding is thought to have accomodated only a limited amount of strain within the shear zone. This fabric represents an intermediate stage in the transition to the more stable, strain-accomodating fabric of Phase Four, shear bands.

Phase 4

Continued deformation resulted in the attenuation and eventual truncation of the Phase Three fold limbs with the development of ductile shear bands (Fig. 15H). The shear bands are very planar, thin, micaceous layers that clearly truncate the preexisting mylonitic fabric, and are oriented at a low angle to that foliation. Some bands have only minor displacement across them, as the Sm schistosity can be traced across them. Thicker bands separate slabs of mylonite that have experienced a greater amount of internal rotation. The concentration of sericite within the shear bands indicates that retrogression of feldspar is important as a softening mechanism, leading to their development, and continued deformation along them. Subsequent deformation resulted in the rotation and fragmentation of the intervening ultramylonitic fabric (Fig. 15, I and J). Ultimately, the rock has the appearance of a breccia composed of elongate clasts displaying a folded ultramylonitic fabric within them (Fig. 15K). The lack of a younger fabric in the shear zone indicates that these shear bands are a late-stage phenomena in the deformational process.

Qz veins and Sericite Phyllonite

Thick sericite phyllonite layers and milky quartz veins are thought to mark zones of extensive fluid flow in the Cobb Hill thrust zone (Fig. 15L). These sericite phyllonite zones are interpreted to result from retrogression of the gneiss during faulting. They are typically strongly foliated, and are concordant with the fabric in the surrounding gneiss suggesting that they formed synchronously with the deformation in the shear zone. The existence of mylonitic gneiss and fragmented gneiss surrounding these zones indicates that they developed during a later stage of the deformation. It is likely that much of the strain within the Cobb Hill thrust zone was concentrated in these phyllonite zones, subsequent to their development, due to the comparatively weak nature of the rock. The presence of thick quartz veins in the shear zone indicates the occurrence of dilatent periods during the fault zone evolution.

EXPLANATION

Paleozoic foliation (thin lines), and Grenvillian schistosity outlined by flattened and elongate quartz grains (black ellipses)

Paleozoic fault zone fabric illustrating the abundance of mica (thin lines), and recrystallized quartz and feldspar (dots)

Quartz Rods (heavy lines)

Feldspar porphyroclasts with asymmetric tails

Shear Sense
Toward
Away

Left lateral

Right lateral

XZ View oriented perpendicular to the fault zone foliation and lineation

YZ View oriented perpendicular to the fault zone foliation and parallel to its associated lineation

Figure 15

Structural Data

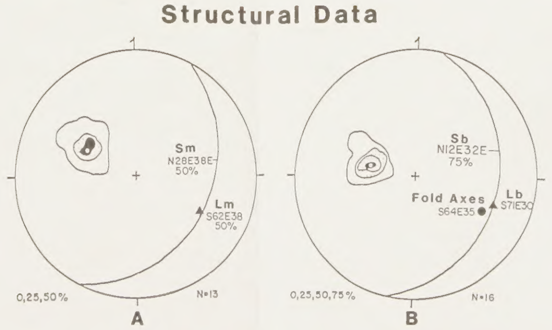


Figure 16A and 16B. A) Lower hemisphere equal area projection of poles to the mylonitic schistosity (Sm) at location V13 (Pl.2) and associated lineation (Lm). B) Lower hemisphere equal area projection of poles to the shear band foliation (Sb) at location V13 (Pl.2), with associated lineation (Lb). Contour interval represents % of data per 1% area. Plane of projection is horizontal. Note that the Sb fabric is more gently dipping than Sm. Both fabrics contain a strong down-dip mineral lineation, and display a consistent east-over-west shear fabric parallel to that lineation. Tick mark denotes north direction. Note that the fold axis lineation of the Phase Four folds is subparallel to Lb, suggesting that they were rotated during shear band development.

Phase 3 - Folded Mylonitic Schistosity

Figure 8. Folded mylonitic schistosity is observed in the same exposure as the Phase Two fabrics previously described viewed looking east. The pervasive mylonitic schistosity (Sm) is well-developed and defined by thin laminae of quartz and feldspar. The Sm schistosity is folded by 1 to 2 cm wavelength folds. The folds have a common north-over-south asymmetry when viewed down-plunge. The hinges of these folds are sub-parallel to the mineral lineation in the rock (see Fig. 16). This is the only exposure along the Cobb Hill thrust zone that preserves the folded Sm schistosity, suggesting that this phase is unstable, marking the transition from a pervasive mylonitic schistosity (Phase Two) to shear band development (Phase Four). Board divisions in cm.

Figure 9. In this hand specimen (above), the folded mylonitic schistosity, Sm, is viewed looking east in the down-plunge direction. Note the few thin micaceous (dark) bands that transect these folds (center of photo). These thin bands develop into a spaced, shear band foliation (Phase Four) which forms as a result of shearing out minor folds along their attenuated limbs. In the lower part of the photo, Sm is only weakly folded.

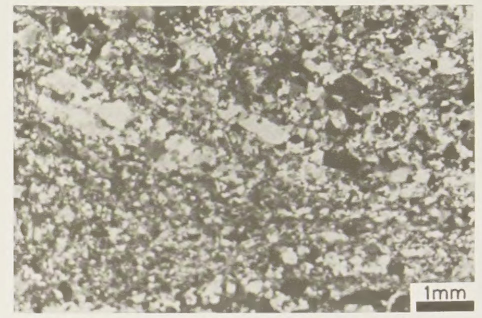


Figure 10. The fine-grained, strongly recrystallized quartz-feldspar-rich mylonitic fabric of the rock is illustrated in the photomicrograph above, viewed parallel to the mineral lineation in the rock. The Sm schistosity is outlined by elongate quartz ribbons (upper left to lower right). Note the bending of these ribbons (center). Sericite is common, but typically not abundant, suggesting that chemical alteration of feldspar was not extensive.

Phase 4 - Shear Bands and Fragmented Mylonite

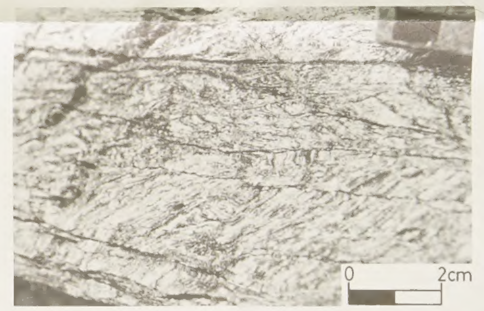


Figure 11. Outcrop photo viewed looking east at Phase Four shear bands. The Phase Four fabric evolves by shearing-out the small-scale Phase Three folds, developing into mica-rich, spaced, shear bands (Sb). These bands clearly truncate and offset the preexisting mylonitic fabric (Sm). Note the planarity of this fabric. Some shear bands appear discontinuous along strike, and are believed to have only minor displacement across them.

Figure 12. Outcrop photo viewed looking south at shear bands. The shear bands develop a C-S relationship with the preexisting mylonitic schistosity (Sm). The C-S fabric and asymmetric, mylonitic gneiss clasts indicate a consistent east-over-west sense of shear in the outcrop when viewed perpendicular to the strong mineral lineation in the rock. Board scale divisions in cm.

Sericite Phyllonite and Quartz Veins

Figure 13. The northernmost of the three outcrops at location V13 (Pl.2) exposes a 5 ft. thick zone of sericite phyllonite and milky vein quartz. The zone (viewed above looking south) is strongly foliated and bounded by mylonitic gneiss. The contact between the gneiss and phyllonite is gradational. Strongly sheared pods of gneiss occur within both the sericite phyllonite and the vein quartz. It is apparent that the gneiss along this zone is strongly sericitized. This zone is indicative of extensive alteration of the gneiss due to the intrusion of hydrous, silica-rich fluids into the fault zone, at some time during its evolution. Board is 13 cm long.

Figure 14. The hand specimen above shows the relationships between the gneiss, phyllonite, and vein quartz viewed perpendicular to the mineral lineation in the rock. Note that the gneiss clasts are completely surrounded by a matrix of nearly-pure sericite. These clasts are strongly elongate parallel to the mineral lineation in the zone. In thin section (not shown), the gneiss clasts appear strongly recrystallized, displaying elongate quartz ribbons, and feldspar grains with a core and mantle texture.